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The distribution of trace elements during strong fractionation of basic magma—a further study of the Skaergaard intrusion, East Greenland

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F. Average values to illustrate the general trend in amounts of major and trace elements during differentiation of the Skaergaard intrusion

ABSTRACT

A number of trace elements have been determined spectrographically in the rocks and minerals of the Skaergaard intrusion, East Greenland. The original basic magma from which the varied rocks of the complex were developed is shown to have had a normal trace element composition. The sorting out of the trace elements into the various mineral series produced by strong fractional crystallization of the original basic magma is traced in detail by means of analyses of the separated minerals. Certain of the trace elements (Cr, Ni) are shown to be strongly concentrated in the early rocks so that later fractions have little or none of them; other elements (P, V, Cu, Sc, Mn, S) reach maximum values in the middle, or late middle stages represented by certain olivine-free gabbros and ferrogabbros; other elements (Li, Zr, Y, La, Ba, Rb) tend to remain in the residual liquid during fractionation and are thus abundant in the latest granite fraction. Still other trace elements (Co, Sr, Ga, Mo) show only small changes in amount throughout the series. Of these Co is a little more abundant in the early and middle stages, Sr in the middle stages, Ga in the later stages and Mo in the early and later but not in the middle stages. The distribution of the trace elements in the rocks is considered in relation to the varying composition of the minerals produced by the fractional crystallization processes and an attempt is made to discuss the mineral compositions in terms of crystal chemical concepts.

The Skaergaard sequence of differentiation from gabbros, through ferrogabbros, to granite is considered to be a common trend of fractionation of basic magma at high levels in the crust, and the observed changes in trace element composition are therefore regarded as having wide geochemical significance. The trace element composition of the intermediate Skaergaard differentiates is significantly different from that of diorites reported by other workers and suggests that diorites have had some other origin than by fractionation of basic magma. On the other hand the trace element composition of many granites resembles that of the granite fraction produced in the Skaergaard intrusion.

I. Introduction

The amounts of trace elements in basic rocks show a certain degree of uniformity, and generally a marked contrast with the amounts in granites. During crystallization of basic magma the trace elements enter the various solid phases in different concentrations dependent, as Goldschmidt has shown, on such factors as their ionic size and charge in comparison with those of the major elements which they replace in the crystal structure. How effectively selection can operate must depend on such factors as the perfection of the crystals forming and the extent to which stirring and diffusion allow equilibrium conditions to be reached between the trace elements in the crystal and in the liquid. In the case of the rock-forming minerals an experimental approach to these problems has not yet been made*, but analyses of rocks and minerals for trace elements has shown broadly what happens during the solidification of magmas under natural conditions. Results of the latter kind based on a further investigation of the Skaergaard intrusion which has already been the subject of a detailed petrological investigation (Wager and Deer, 1939)† form the subject of this paper. The Skaergaard complex is worth renewed study in this way because it shows particularly clearly the effect of strong crystal fractionation.

A few years ago we presented a preliminary paper on the distribution of the trace elements in the rocks of the Skaergaard intrusion (Wager and Mitchell, 1943). This gave a broad indication of the trends in the amounts of the trace constituents in the rock series. In discussing the results we could not do more than surmise in which minerals the trace constituents would be likely to be located, using earlier analytical data and general crystal-chemical principles enunciated by Goldschmidt. A later investigation was promised of the actual trace element composition of the individual minerals forming the various rocks. The present paper embodies these further studies and also provides additional and more reliable data on the trace element composition of the rocks themselves.

The original paper on the Skaergaard intrusion (1939) contained many analyses of rocks and minerals by W. A. Deer, one of the authors. Dr. Deer has supple-

* Work of this kind has been begun by Holland and Kulp (1949, p. 59).

† The paper by Wager and Deer on the petrology of the Skaergaard intrusion will be referred to frequently and will be cited in the following pages by giving the year of publication only.

mented these results by further chemical analysis of certain minerals and for this we wish to thank him warmly. Mr. E. A. Vincent has also kindly furthered the work by analyses of the sulphur content of certain rocks and Dr. H. Neumann has given us help and advice on the preliminary investigation of certain opaque minerals.

The spectrographic work has been carried out at the Macaulay Institute for Soil Research by one of us (R.L.M.). We wish to express our grateful thanks to the present Director of the Macaulay Institute, Dr. D. N. McArthur, and the former Director, Sir William Ogg, for their enlightened policy of permitting fundamental work in the related field of geochemistry to be undertaken at the Macaulay Institute. An understanding of the distribution of the trace elements in the rocks and minerals of the earth's crust is clearly of great importance to the soil scientist in view of the importance of trace element deficiencies and excesses for the health of plants and animals.

II. Previous Work on the Skaergaard Intrusion

The Skaergaard intrusion forms an area of 60 km^2 of fjord, mountain and glacier on the east side of Kangerdlugssuaq in East Greenland, Lat. 68° 10′ N, Long. 31° 40′ W. Mapping and collection of the rocks of the intrusion were begun in 1930 and continued in 1932 and 1935–36, the field investigations being made by the first-named author (L.R.W.) on the first two occasions, and by him and W. A. Deer on the third and longest period of work in Greenland. Accounts of these expeditions and of the general geology are available (Wager, 1934, 1937 and 1947), while the petrology of the intrusion has been presented in detail in the paper cited above by Wager and Deer (1939). The rocks and minerals of which the trace constituent composition is here investigated have for the most part been described and chemically analysed in the course of the earlier work. The rock specimens, referred to by the collection numbers, are housed in the Department of Geology and Mineralogy, Oxford.

In the original description of the Skaergaard complex it was shown that the complex was at first a pool of basic magma of about 300 km^3 in volume and with the form of an inverted cone which, at the present level of denudation, is 11 km from north to south and 7 km from east to west (Figure 1). During cooling various processes caused marked differentiation, giving rocks which have been grouped as the layered series and the marginal and upper border groups (inset, Figure 2). The names used for the rock types developed during differentiation, and their spacial relations, are given diagrammatically in the main part of Figure 2 together with the collection numbers of the chief rocks for which results are given in this paper.

The lowest layered rocks exposed are hypersthene olivine gabbros and these, when traced upwards, give place gradually to rocks of lower solidification temperatures—first to Middle Gabbros, then to ferrogabbros in which the plagioclase becomes andesine and the olivine and pyroxenes become progressively more iron-rich until they reach fayalite and hedenbergite respectively. Various lines of evidence show that the layered series was formed as a precipitate of discrete crystals, separated from the overlying liquid and accumulated as a sediment from below upwards. The precipitate held some liquid in the interstices at the time of its formation and this necessarily had the composition of the magma existing at the particular stage. As cooling progressed the interstitial liquid crystallized to form outer zones of lower melting point solid-solutions round the crystals of the primary precipitate and, in certain cases, additional crystal phases such as apatite and quartz were formed. The nature and general composition of the chief crystal phases present in the successive layered rocks are shown in Figure 3.

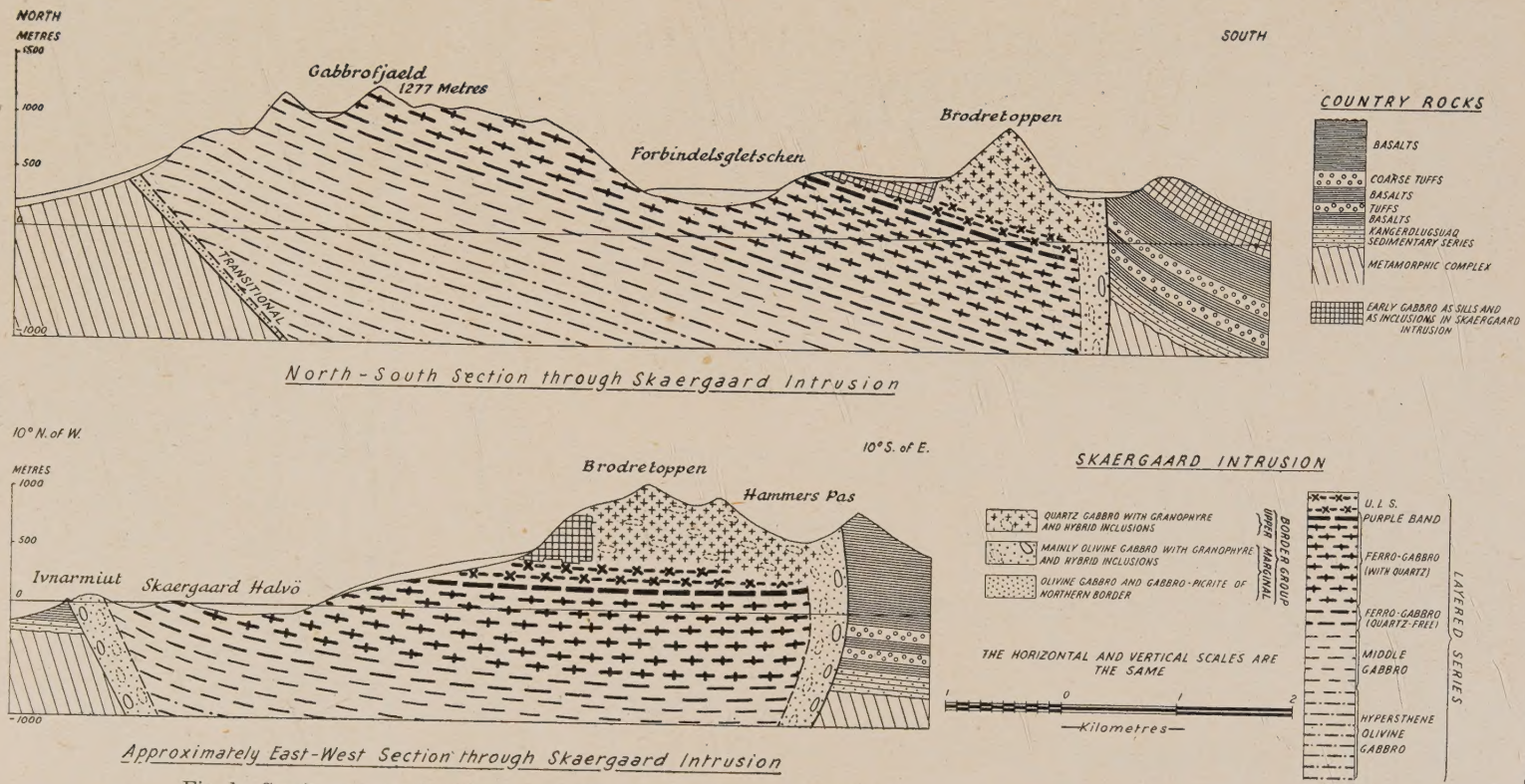


Fig. 1—Sections across the Skaergaard intrusion (reprinted with permission from Meddel. om Grönland, 1939 **105**).

The layered series is characterized throughout by rhythmic layering, the individual layers often showing gravity stratification. Normally the rocks chosen for analysis for trace elements were average material for the particular horizon, so far as could be judged by eye. The modal composition of the average rocks is also given in Figure 3. Two strongly contrasted adjacent bands (2568, 2569) of the layered series have also been analysed to indicate the extent of the effects the banding may have on the trace element composition.

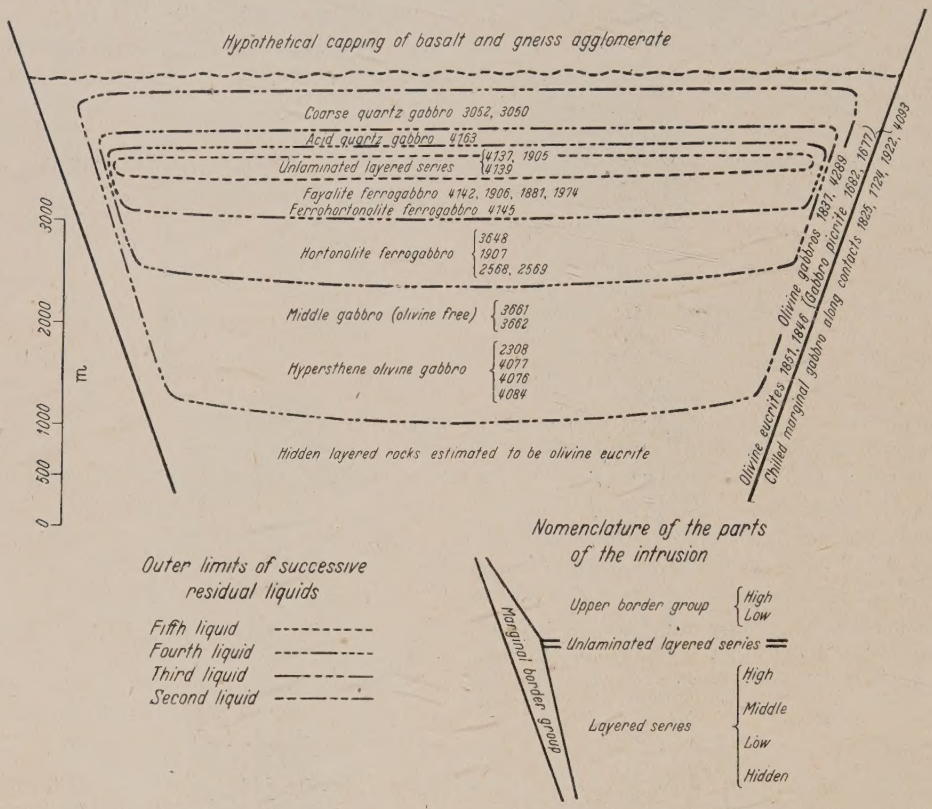

Hypothetical capping of basalt and gneiss agglomerate
Coarse quartz gabbro 3052, 3050
Acid quartz gabbro 4163
Unlaminated layered series 4137, 1905 4139
Fayalite ferrogabbro 4142, 1906, 1881, 1974
Ferrohortonolite ferrogabbro 4145
Hortonolite ferrogabbro 3648 1907 2568, 2569
Middle gabbro (olivine free) 3661 3662
Hypersthene olivine gabbro 2308 4077 4076 4084
Hidden layered rocks estimated to be olivine eucrite
Olivine eucrites 1851, 1846 (Gabbro picrite 1682, 1677)
Olivine gabbros 1831, 4289
Chilled marginal gabbro along contacts 1825, 1724, 1922, 4083
3000
2000
1000
500
0
m.
Outer limits of successive residual liquids
Fifth liquid
Fourth liquid
Third liquid
Second liquid
Nomenclature of the parts of the intrusion
Upper border group High Low
Unlaminated layered series
Marginal border group
Layered series High Middle Low Hidden

Fig. 2—Diagrammatic east–west section across the Skaergaard intrusion showing successive stages in the solidification and the names given to the chief rock units. The numbers refer to specimens considered in the paper and the arrangement shows their relative position.

The layered rocks were formed by a process of crystal fractionation. Thus the composition of any particular rock does not normally represent the composition of the liquid from which it was formed. The original Skaergaard magma is taken to have had the composition of the fine-grained chilled marginal gabbro and this is known from two chemical, and four spectrographic analyses. Gradually the initial composition of the liquid became modified—slowly at first, then more quickly—by the precipitation of crystal phases of steadily varying composition. Certain arbitrary stages in the successive residual liquids are defined in Figure 2. For the major chemical constituents the composition of the liquid at these stages was estimated previously (1939, pp. 218–219) and is estimated for the trace elements in this paper.

The formation of the layered series seems to have dominated the differentiation process, but at the same time there was solidification of materials to give the marginal and upper border groups. The marginal border group, when traced from the margin inwards, varies from rocks of high to low solidification temperatures, much as the layered series varies from below upwards. The marginal border rocks, however, have both compositional and textural differences from the layered series, and they are regarded as largely produced by the solidification of successive

residual magmas, in which discrete crystals, like those giving rise to the layered series, were sometimes incorporated. Contamination of the border group with acid material derived from blocks of the country rock incorporated in the intrusion, complicates the interpretation of the border group and produces contaminated rock types. Of particular interest are olivine eucrites such as 1851 and 1846 which form the outer zone of the marginal border group, and which are believed to be fairly similar in composition to the hidden, lower part of the layered series. In the northern border group a gabbro-picrite is found (1682, 1676) which is an early border rock produced by concentration of early olivine crystals.

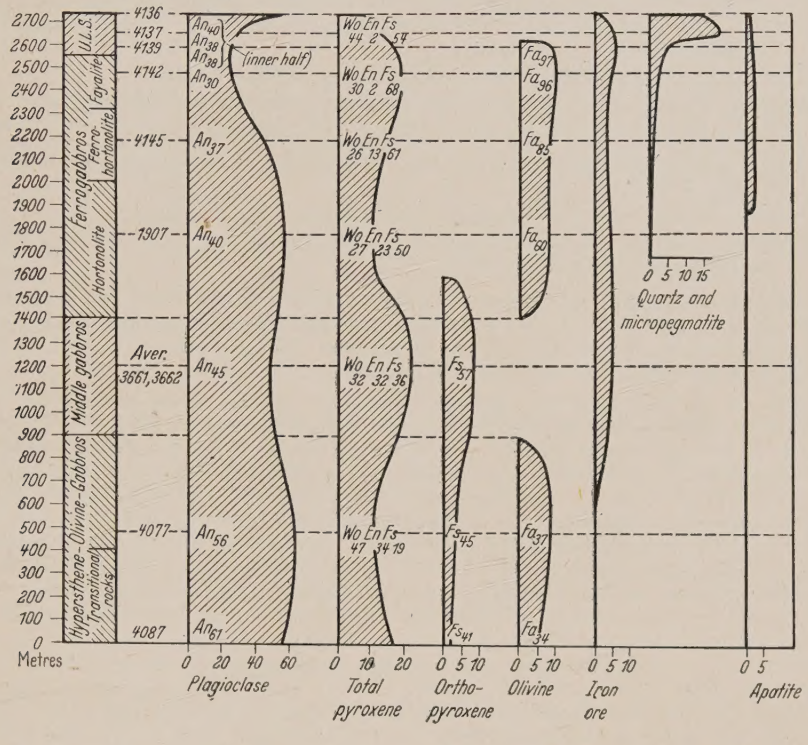


Fig. 3—Modal composition of average rocks of the layered series and the composition of certain of the minerals (reproduced with permission from Meddel. om Grönland 1939 **105**).

The upper border group is heavily contaminated by acid gneiss, presumably because the lower density of the acid material allowed it to float up and accumulate at the top of the complex. Only small remnants of this part of the complex are preserved (Figure 1) but they are sufficient to show what it was like. The highest upper border rocks (3050, 3052) are interpreted as having been formed from the magma in existence at the time of formation of the Middle Gabbros and contaminated with acid gneiss from the country rock. The low horizons of the upper border rocks, such as 4163, are interpreted as the magma of the hortonolite ferrogabbro stage contaminated by acid gneiss.

Towards the end of the solidification processes there was only a shallow sheet of liquid left in the upper middle part of the complex. This separated into a lower layer of fayalite ferrogabbro of special type (4139) and a basic hedenbergite granophyre (4136, 1905). Filter press action apparently occurred at about this stage and produced indefinite sill-like sheets of hedenbergite granophyre and later still, a sill of acid granophyre, together with irregular veins and dyke-like masses.

Although the processes of differentiation were continuous, it is convenient to divide the various rocks produced into arbitrary stages as follows:

A Gabbro picrites of the northern border group.

A^1 Earliest olivine gabbros of the outer part of the eastern and western border groups which were formed at about the same stage as the gabbro picrite.

A^2 Olivine gabbros of the main part of the eastern and western border groups, probably roughly contemporaneous with the earliest exposed layered rocks.

B Hypersthene olivine gabbros—the lowest exposed layered rocks—occurring from 0–900 m in the layered series on the arbitrary height scale.

C Olivine free gabbros (the " Middle Gabbros ") occurring from 900–1400 m in the layered series.

D Hortonolite ferrogabbros, 1400–2000 m in the layered series.

E Ferrohortonolite ferrogabbros, 2000–2300 m in the layered series.

F Fayalite ferrogabbros, 2300–2550 m in the layered series.

G Basic hedenbergite granophyres, the latest rocks of the unlaminated layered series (SiO_2 *C.* 52%).

H Certain transgressive hedenbergite granophyres (SiO_2 *C.* 58%).

I Acid granophyres, the latest rocks of the Skaergaard intrusion, forming thin sills and irregular dykes (SiO_2 *C.* 75%).

In the tables accompanying this paper the various stages listed above are indicated by the letters used in the above list; in the same way the values for the average Skaergaard magma are always indicated by Y, and the values for average " Grey Gneiss ", the dominant rock of the metamorphic complex, are indicated by Z.

The rocks of the complex were formed from the original magma, together with some extraneous material from the country rock which became incorporated with it. The effect of some degree of contamination by acid gneiss can be seen in certain granophyre and hybrid masses of the border group and the lower layered series. After the hortonolite ferrogabbro stage no patches of acid granophyre are found and it is considered that the magma had by then become homogeneous. Although the magma must have been modified to some extent by addition of material from the xenoliths which passed completely into solution, all variation in the rocks produced subsequently to the hortonolite ferrogabbro stage is regarded as solely the result of differentiation of the homogeneous and slightly contaminated magma. Thus the hedenbergite and acid granophyres discussed in this paper are held to be the result of solidification of residual liquids produced by fractionation. Certain small veins in the high upper border-group and the outer marginal border group which can be traced to neighbouring granophyre masses have also been described (1939, pp. 20, 23–24); these are interpreted as material from the re-melting of gneiss inclusions which locally injected the surrounding solid, but still hot, gabbro. They are to be distinguished from the more extensive sills, veins and dykes of acid granophyre produced by filter press action during the late stages of fractionation; only the latter have been analysed for trace elements.

The rock types developed by fractional crystallization of the original rather felspar-rich, olivine gabbro magma were, in the early stages, fairly normal eucrites and gabbros. Then came the more iron-rich Middle Gabbros without olivine, and still later, the unusual iron-rich rocks to which the name ferrogabbros was given in the original memoir. The final rocks, however, are of acid granophyre composition and are fairly similar to normal soda-granites.

III. Methods Used and the Accuracy of the Data

1. Preparation of the rocks and minerals for analysis

For the spectrographic analysis of the rocks, specimens weighing 20–30 g were used, but somewhat larger amounts were taken of the coarser rocks when the available material was sufficiently abundant. Had the region been more accessible rather larger samples would have been used to ensure their being fully representative.

The minerals which have been analysed are from fairly coarse rocks for the most part, and a sufficient amount of pure material was obtained without undue labour by crushing, sieving to a proper size, and hand picking. The crushing was all done in one particular small steel mortar. Silk fabrics were used for sieving and not brass sieves. The size selected for hand

picking varied with the different minerals; it depended on the size required to give essentially mono-mineralic grains and also on the degree of transparency of the minerals. The hand picking for all except the iron ores was done in transmitted light as this gave the best test of homogeneity. For iron-rich pyroxenes and olivines it was necessary to crush more finely than in the case of the more magnesium-rich, because of their dark colour. Separation by heavy liquids or by magnetic methods was usually unnecessary and was not used except in obtaining the apatites and iron-ores. The specimens of iron-ore used in the spectrographic analysis were no doubt more contaminated than is the case with the non-opaque minerals, a point dealt with when the iron-ores are considered. The hand picking was done under a binocular microscope and between 0·02 and 0·05 g of the mineral was used for the spectrographic work.

2. *The spectrographic method used*

The method employed for the spectrographic analysis of the rocks and minerals has been fully described elsewhere (Mitchell, 1940; 1948). The finely ground sample, mixed with an equal weight of carbon powder was burned in a carbon electrode, in a 9 amp DC arc, with the material in the cathode. Spectrograms covering the wavelength range 2500Å to 8500Å were compared in a *Hilger Judd Lewis Comparator* with those prepared in a like manner from standard mixtures with a matrix similar to that of the material being examined.

By this method it is possible, with practice and over favourable density ranges of the spectral lines, to obtain an accuracy of $\pm 30\%$ when samples and standards are closely similar in basic composition. It was not possible to prepare standards similar in composition to all the materials examined, so somewhat greater errors may have occurred in such instances, *e.g.* for certain of the mineral species, and also where the contents of the trace elements exceeded a value of 1000 ppm. For many elements, including cobalt, nickel, vanadium, chromium, copper, strontium, barium and lithium this accuracy would appear to be obtainable over the range of 10–500 ppm. For zirconium, and probably also for the rare earths, high values become uncertain in the presence of silicon unless very stringent precautions are taken. In general, however, in our discussion we have considered $\pm 50\%$ to be a reasonable assessment of the accuracy obtained and have considered differences greater than this to be significant. It is not claimed that the values are very precise; the semi-quantitative method has the advantage of speed, allowing many more samples to be examined, for many more elements, than would be possible by a truly quantitative method. The variations in amount of most constituents are much greater than $\pm 50\%$. In addition in considering the value of the analytical method employed the magnitude of possible sampling errors must be taken into account.

3. *Discussion of the accuracy of the data*

The spectrographic method employed, as explained in the previous section, gives values considered to be within half and twice the true value, although usually much closer. The order in which the samples were spectrographically analysed was haphazard and no trend, therefore, was in the mind of the observer when reading the plates. When the data were assembled it was generally a matter of satisfaction to find how well they fell into position in the various series of results.

The fragment of the hand specimen taken as a sample for analysis was, no doubt, normally large enough to be representative. The rocks, however, were formed by the accumulation of discrete crystals and when seen in large masses in the field they usually showed banding. Average material from among the banded rocks was selected in the field by eye only. The work on the major constituents (1939) showed that on the whole the material analysed was reasonably representative of its particular horizon in the layered series, with the exception of the Middle Gabbros where an average was obtained by analysing two neighbouring rocks, 3661 and 3662. From point of view of the trace constituents it is considered that the rocks analysed, with few exceptions, represent reasonably well the average material for the particular stage of differentiation. The exceptions are for certain trace elements like zirconium; at a late stage this element is present in discrete zircon crystals which are so thinly scattered through the rock that our samples may not always have been large enough to give a fair value.

In considering the data for the trace constituents of the separated minerals the extent of contamination of the sample by other minerals of the rock has to be considered. An estimate of the maximum possible contamination by any of the other analysed minerals of the rock is often to be obtained from consideration of the whole data on any particular rock. Thus con-

sidering the minerals from the hypersthene-olivine gabbro 4077 (Table 1) the amount of chromium in the pyroxene is on the average 350 ppm and, therefore, one part of pyroxene in 350 parts of felspar would bring the amount of chromium above the sensitivity, which is 1 ppm. However, the analysis shows that chromium in the felspar is below 1 ppm and, therefore, that the felspar sample has less than 0·3% of pyroxene as impurity. This method may frequently be used to obtain the maximum possible contamination by other analysed minerals of the rock but it cannot of course be applied if contamination results from small amounts of some mineral whose composition has not been determined as, for instance, hornblende or chlorite formed by late stage reactions.

Besides contamination by other major mineral constituents of the rock there are minute inclusions of various sorts scattered through the crystals and also orientated crystallites which may be due to ex-solution. These are described in the appendix where their effect on the mineral analyses is also considered.

During the original chemical analysis for the major constituents no systematic determination of trace elements was attempted, but in five cases the chemical analysis was extended to include determination of Cr, Ni, Cu, Zr, Sr and Ba (1939 Table XLI). The correspondence between the chemical and spectrographic data in the case of chromium is good when " trace " and " nil " of the chemical analysis are interpreted as 200 ppm or less. Nickel shows poor correspondence; copper, except in one case shows good correspondence; zirconium shows reasonable correspondence when " nil " is interpreted as 50 ppm or less. Strontium and barium show from 2–3 times more by chemical analysis and it seems probable that the chemical determinations of these elements were unsatisfactory. The spectrographic determinations of strontium and barium also proved difficult, the figures now given being considerably less than those in the preliminary paper (Wager and Mitchell, 1943).

Comparison with such chemical data as are available forces the attention on the errors that may occur in the absolute amounts. While believing that our present spectrographic results are within the limits of error discussed in the previous section, it is important to realize that the relative values for the whole data are consistent and that relative values are sufficient for many of the deductions reached.

As a general test of the reproducibility of our results the determinations on the hypersthene olivine gabbro 4077 and its component minerals were repeated. The rock analyses were carried out on two different occasions on two separate fragments broken from the single hand specimen. The mineral analyses were carried out on two different occasions from two sets of separately prepared samples of the crushed rock. The results which are presented in Table 1 are usually well within the limits of error of the spectrographic technique except that the nickel of the pyroxene is on the verge of these limits. This is probably accidental as determination of nickel is straightforward and usually gives good results. Re-examination of the spectrograms suggests that the two values for nickel should actually be closer, but we have retained the first readings. In this case, and in all others, the readings were not revised in the light of any conceptions obtained when the data were assembled, although in a few cases, which are always mentioned, further analyses were made of the same or related material, and averages taken.

IV. Composition of the Original Skaergaard Magma

The fine-grained olivine gabbro, found at all outer contacts of the intrusion is taken to be the quickly chilled, original magma. Chemical analyses were made of two different specimens (1825 and 1724) and now these two specimens and two others (1922 and 4093) have been analysed for trace elements (Table A)*.

In this and all other tables the trace elements are given in order of size of their ions at the usual valency in igneous minerals. The ionic sizes are those given by Evans (1939, p. 170). In our first paper (Wager and Mitchell, 1943) this plan was adopted for all the elements dealt with but in this paper we have put the major elements (see p. 178 for definition) in the first part of the table as has been done by Nockolds and Mitchell (1948).

* The tables described by the letters A–F will be found at the end of the paper. They contain most of the new spectrographic data. The other tables in the body of the paper, numbers 1–15, are largely comparative but in some cases include additional new determinations.

Table 1—Full spectrographic data for hypersthene-olivine gabbro 4077 at 500m in the layered series and for plagioclase, pyroxene and olivine separated from it

(Figures are in parts per million)

	r	s	*Rock*			*Plagioclase (An56)*			*Pyroxene (Wo47 En34 Fs19)*			*Olivine (Fo64)*			
			A	*B*	*Average*	*A*	*B*	*Average of A & B*	*A*	*B*	*Average*	*A*	*B*	*Average*	
P	0·34	—	—	—	(260)	—	—	—	—	—	—	—	—	—	P
Ga	0·62	1	20	25	**23**	60	40	**50**	*	5	**3**	5	*	**3**	Ga
Cr	0·64	1	150	200	**175**	*	*	*	400	300	**350**	*	*	*	Cr
V	0·65	5	250	200	**225**	*	*	*	200	300	**250**	*	*	*	V
Mo	0·68	1	*	*	*	*	*	*	2	3	**3**	10	10	**10**	Mo
Li	0·78	1	2	2	**2**	2	4	**3**	*	3	**2**	3	3	**3**	Li
Ni	0·78	2	150	120	**135**	*	*	*	80	200	**140**	400	250	**325**	Ni
Co	0·82	2	60	50	**55**	*	*	*	40	60	**50**	100	150	**125**	Co
Cu	C.0·8[1]	10	100	60	**80**	50	20	**35**	30	40	**35**	20	20	**20**	Cu
Sc	0·83	10	20	20	**20**	*	*	*	30	30	**30**	*	*	*	Sc
Zr	0·87	10	40	30	**35**	*	*	*	40	60	**50**	*	*	*	Zr
Mn	0·91	10	—	—	(700)	—	20	**C. 20**	—	—	(2100)	—	—	(1700)	Mn
Y	1·06	30	*	*	*	*	*	*	*	*	*	*	*	*	Y
La	1·22	30	*	*	*	*	*	*	*	*	*	*	*	*	La
Sr	1·27	10	700	700	**700**	1000	1000	**1000**	*	*	*	10	10	**10**	Sr
Ba	1·43	5	20	30	**25**	80	50	**65**	10	*	**7**	10	*	**7**	Ba
Rb	1·49	20	*	*	*	*	*	*	*	*	*	*	*	*	Rb

NOTES:

A and B results were determined on different samples of the same handspecimen and on different samples of the minerals separated on different occasions.

The major constituents of the rock are given in Table C column VIIB and the major constituent of the minerals in Table B columns II, VIII and XIV.

r is ionic radius in Ångstrom units at valency shown.

s is sensitivity with spectrographic method employed.

Data was obtained spectrographically except where figures are in brackets when the values are from chemical analysis, *e.g.*, P and Mn.

An asterisk indicates that the element was present in amounts less than the sensitivity.

A dash indicates absence of data.

The following elements were also sought but are not present in amounts above the sensitivity shown in brackets: B(1), Be(10), Ge(10), Sn(5), Zn(300), Sb(300), Ce(500), Cd(300), Ag(1), Pb(10), Tl(50).

The pyroxene is dominantly clino-pyroxene of the composition shown but a little ortho-pyroxene is also present.

[1] See GOLDSCHMIDT 1946.

A description of the chilled marginal rocks, which in appearance are normal olivine gabbros, has previously been given (1939, p. 137). The major element composition shows a close resemblance to the Mull Porphyritic Central Magma Type of the appropriate silica percentage (Bailey, *etc*, 1924, pp. 14–24). A still closer approach is obtained by averaging two parts of the Porphyritic Central Magma Type, SiO_2 48% and one part of Non-Porphyritic Central Magma Type, SiO_2 47% and it is suggested that the Skaergaard magma belongs to a type of basic magma widely available in the upper crust. There are, however, two ways in which the Skaergaard magma differs from the weighted average of the two Mull magmas: first, the iron of the Skaergaard magma is more reduced and, second, the amount of potassium in the Skaergaard magma is only one third of the average. These differences are regarded as significant but they do not make it necessary to consider the Skaergaard magma as in any way unique since gabbros and basalts almost as reduced as the Skaergaard magma and having similarly low potassium are known (see for instance average data assembled by Walker and Poldervaart, 1949, p. 349).

The trace constituents of these four rocks (Table A) show some variation which is outside the limits to be expected from the spectrographic technique; for instance, the variation in some cases is considerably greater than that shown by the two analyses of the same specimen of hypersthene-olivine gabbro 4077 (Table 1). Some variation is to be expected as the specimens were taken at varying distances from the margin and slight differentiation may well have taken place. In striking an average for the original Skaergaard magma (column Y) the data for 4093 have been omitted as there seems to have been enrichment in some chromium bearing material, probably chromite. This particular specimen comes from the northern border group where, at a later stage, gabbro picrite was developed which is also a rock showing enrichment in chromium (Table C). The amounts of the trace elements in the chilled marginal gabbros, and, therefore, by inference in the original Skaergaard magma, show the following features:—V, Ni and Cu between 100 and 200 ppm, Co and Zr around 50 ppm; Sr about 350 ppm; Ba about 50 ppm, and Ga and Sc between 10 and 20 ppm.

It is interesting to compare this trace element composition with such other data as are available. For comparison purposes we have determined the trace constituents of three basalts from east Greenland (Table A, columns V, VI and VII) previously analysed chemically by H. F. Harwood (Wager, 1934, pp. 32–33). All three are olivine basalts but two (1093 and 1112) carry more olivine and have an average trace element composition close to that of the Skaergaard magma (*cf* columns Y and M). The third basalt (1057) is a tholeiitic type with only a small amount of olivine and it is considered to have come from a more differentiated and perhaps contaminated magma. In harmony with this, the amounts of chromium and nickel are much reduced compared with the other two basalts, and copper, zirconium and barium are somewhat increased. These few results suggest that the trace element composition of the original Skaergaard magma is essentially similar to that of the common olivine basalts of the area. The volume of magma in the Skaergaard intrusion is large, and the intrusion was emplaced quickly. The magma forming the intrusion must have been abundantly available locally within the earth. Likewise the basalts of the region are of great volume and must represent an abundantly available magma. The general resemblance between the trace element composition of the Skaergaard magma and of the common type of East Greenland

basalt suggests that the Skaergaard magma must be regarded as belonging to an abundant and widespread magma type of the East Greenland area.

Comparison with other data for basic rocks (columns N to Q) shows that the Skaergaard magma had a general trace element composition like that of other regions although the amount of data with which comparison can usefully be made is limited. The figures for a composite sample of eleven gabbros (six classified as gabbros, four as olivine gabbros, and one as norite) which was prepared by NOLL and analysed by various workers, provide the most interesting comparison and this was indeed the only comparison we made in our original paper (1943, p. 285)*. The figures available (column O) show a good correspondence with our average figures (column Y) except in the chromium, but here the possibility of concentrations of chromite in some rocks and not in others reduces the significance of variations in chromium content. Certain average data (column P) from SANDELL and GOLDICH (1943) based on graphs which are the results of their many chemical determinations, again show good agreement with the Skaergaard averages. Data from two Karroo dolerites analysed by MITCHELL (WALKER and POLDERVAART, 1949, p. 644) are also given (column N) because they are representatives of a widespread basic magma type; on the whole they too are reasonably similar to the Skaergaard magma. Finally some averages calculated by TRÖGER (1935, p. 316) and based on ordinary chemical analyses are given (column Q). These are of interest where more recent determinations are not available, as for instance in the case of phosphorus, vanadium, manganese and sulphur. Further comparisons are made when the separate trace elements are considered in Section X, but it is clear from those already given that the Skaergaard magma had a reasonably standard trace element composition.

V. Trace Elements in Certain Selected Rocks and in Minerals Separated from them

The manner in which the trace constituents of a rock are distributed among the various crystal phases forming it is well shown by our determinations. From many cases three will be considered in detail: the hypersthene olivine gabbro 4077, the hortonolite ferrogabbro 1907 and the chilled marginal rock 1724.

The hypersthene olivine gabbro and the hortonolite ferrogabbro are units of the layered series, formed from a precipitate of discrete crystals together with the products of crystallization of the interprecipitate magma. In the case of the hypersthene-olivine gabbro the primary precipitate minerals are plagioclase, clino-pyroxene and olivine, and the chief additional phases from the interprecipitate magma are ortho-pyroxene, iron ores and apatite. For the hortonolite ferrogabbro the primary precipitate minerals are plagioclase, clino-pyroxene, olivine, ilmenite and magnetite, while the additional crystal phases from the interprecipitate magma are quartz and apatite. Only the primary precipitate minerals have been separated and analysed. They are essentially unzoned, their major chemical composition is known in most cases and, by hand picking, a product of a high degree of purity has been obtained.

The way in which the trace elements enter the three primary precipitate crystal phases of the hypersthene-olivine gabbro 4077 are shown in Table 2. Features of interest are given below.

* In the previous comparison of the Skaergaard magma with NOLL'S composite gabbro (WAGER and MITCHELL, 1943, pp. 285 and 288) the high value for strontium in the Skaergaard rock was particularly commented upon. It is now clear that these first spectrographic determinations of strontium from the Skaergaard rocks were considerably too high. There is no evidence from the present revised data for suggesting that this region of the crust might be abnormally rich in strontium.

Table 2—The major and trace constituents of the hypersthene-olivine-gabbro, 4077 and of its individual minerals.

	r	s	*Plag.* An 56 (48% *in rock*)	*Pyrox.* En34 (28½% *in rock*)	*Oliv.* Fo64 (21½% *in rock*)	*Rock*	Σ *for minerals except iron ores*	*Amounts in* 48% *Plag.*	*Amounts in* 28½% *Pyrox.*	*Amounts in* 21½% *oliv.*	*Estimated composition of magma*	
			Major constituents as weight per cent of the oxide									
SiO_2	0·39		54·35	49·89	38·11	**46·37**	**48**	26·0	14·2	8·2	**47**	SiO_2
Al_2O_3	0·57		28·35	3·74	nil	**16·82**	**14·5**	13·5	1·0	—	**15**	Al_2O_3
TiO_2	0·64		nil	0·93	trace	**0·79**	**0·25**	—	0·25	—	**2**	TiO_2
Fe_2O_3	0·67		0·49	0·21	0·15	**1·52**	**0·3**	0·24	0·06	—	**3**	Fe_2O_3
MgO	0·78		0·02	14·59	30·50	**9·61**	**10·8**	—	4·2	6·6	**6**	MgO
FeO	0·83		0·18	12·89	31·48	**10·44**	**10·5**	—	3·7	6·8	**14**	FeO
Na_2O	0·98		4·95	0·43	—	**2·45**	**2·5**	2·4	0·1	—	**3**	Na_2O
CaO	1·06		11·19	16·83	0·02	**11·29**	**10·2**	5·4	4·8	—	**10**	CaO
K_2O	1·33		0·38	0·13	—	**0·20**	**0·21**	0·18	0·03	—	**0·3**	K_2O
			Trace elements as ppm of the element (values are averages of two determinations, see Table 1)									
P	0·34		—	—	—	**(260)**	—	—	—	—	**1500**	P
Ga	0·62	1	50	3	3	**23**	**24**	24	—	—	**20**	Ga
Cr	0·64	1	*	350	*	**175**	**100**	—	100	—	**50**	Cr
V	0·65	5	*	250	*	**225**	**71**	—	71	—	**170**	V
Mo	0·68	1	*	3	10	*	**2**	—	—	2		Mo
Li	0·78	1	3	2	3	**2**	**2**	1	0.5	0.5	**4**	Li
Ni	0·78	2	*	140	325	**135**	**110**	—	40	70	**35**	Ni
Co	0·82	2	*	50	125	**55**	**41**	—	14	27	**30**	Co
Cu		10	35	35	20	**86**	**31**	17	10	4	**245**	Cu
Sc	0·83	10	*	30	*	**20**	**9**	—	9	—	**13**	Sc
Zr	0·89	10	*	50	*	**35**	**14**	—	14	—	**90**	Zr
Mn	0·91	1	c.20	(2100)	(1700)	**(700)**	**(980)**	c.10	(600)	(370)	**(1250)**	Mn
Y	1·06	30	*	*	*	*	—	—	—	—	**20**	Y
La	1·22	30	*	*	*	*	—	—	—	—	**12**	La
Sr	1·27	10	1000	*	10	**700**	**480**	480	—	2	**495**	Sr
Ba	1·43	5	65	7	7	**25**	**34**	31	2	1	**100**	Ba
Rb	1·49	2	*	*	*	*	—	—	—	—	**5**	Rb

An asterisk indicates that the element is present in amounts less than the sensitivity.
The percentages given for the minerals in the rocks are weight percentages.
r is the ionic radius in Angstrom units at valency shown.
s is sensitivity with spectrographic method employed.

Gallium enters the plagioclase abundantly and is very low in the pyroxene and olivine. The amounts in the pyroxene and olivine are not significantly different. (This is also true in other cases discussed later. As the gallium is presumably replacing aluminium it was anticipated that the amount in pyroxene, where there is 3·7% Al_2O_3, would be markedly higher than the amount in the olivine.)

Chromium is less than the sensitivity (*i.e.* less than 1 ppm) in the plagioclase and olivine, while it is 350 ppm in the pyroxene. The way in which Cr readily enters the pyroxene and shuns the olivine is clearly shown.

Vanadium also enters pyroxene but not olivine. It may enter plagioclase in amounts near the sensitivity value; results from other plagioclases suggest this, but re-examination of the spectrographic plates for the two analyses of this plagioclase has confirmed that it is doubtful whether V is present or not.

Molybdenum is present in both pyroxene and olivine but is definitely more abundant in the latter—a result which was not expected.

Lithium is present in small and about equal amounts in all three crystal phases.

Nickel and *Cobalt* are present abundantly in olivine and are about half as abundant in the pyroxene, while these elements are below the sensitivity (less than 2 ppm) in the plagioclase.

Copper is present about equally in all three crystal phases.

Scandium and *Zirconium* are present in the clino-pyroxene and are below the sensitivity in the plagioclase and olivine.

Manganese is abundant in the pyroxene and olivine in which minerals it was determined chemically. It is present in small amount (20 ppm) in the plagioclase where it was determined spectrographically.

Strontium is abundant in the plagioclase being about 1000 ppm (chemical determination gave SrO as 0·24%, *i.e.* Sr=2000 ppm) and it is low in pyroxene and olivine. Its presence in the pyroxene replacing calcium might have been expected, but actually it is below the sensitivity (10 ppm) while in the olivine it is just above. It is probable that the strontium in both the pyroxene and olivine is very low and that the amounts recorded are largely due to contamination of the samples by plagioclase.

Barium is present in the plagioclase, but in much smaller amount than strontium, and it is round about the limit of sensitivity (5 ppm) in pyroxene and olivine.

From the figures presented it is abundantly clear that when crystals are formed by a clean process of fractional crystallization, as in this case, they exercise a strong selective action on the trace elements present in the magma. It is also clear that pyroxene provides a suitable environment for a greater number of ions than either plagioclase or olivine; in particular it receives the various tri- and tetra-valent ions such as Cr, V, Sc and Zr. This has been known in a general way from previous analyses but the data given here provide perhaps the most satisfactory demonstration.

The primary precipitate forms the bulk of the layered rocks—probably 80–90% (unfortunately there seems to be no way of measuring the proportion accurately). For a given trace element the contribution which the various primary precipitate crystals should make towards the total amount of the trace element in the rock has been calculated (second half of Table 2) using the figures for the modal* percentages of the plagioclase, pyroxene and olivine. A comparison of the trace element composition of the rock with the summation of the contributions from the three primary precipitate crystals shows that the values are of the same order. This should be the case as the three primary precipitate crystals make up the bulk of the rock. If the results were sufficiently accurate the discrepancy between the rock composition and the summation of the contributions due to the primary precipitate minerals should give the composition of the interprecipitate magma. The composition of this magma which is estimated from the available data in a later section of this paper (pp. 174–5) is quoted here for consideration in relation to the present data. The figures for the trace constituents in the rock should differ from the summations for the primary precipitate minerals in the same direction as the composition of the contemporaneous magma since the rock consists of the primary precipitate plus the interprecipitate magma. In the case of vanadium, copper and zirconium, where the interprecipitate liquid is estimated to have been markedly richer in these elements than the rock, the expected effect is apparently shown. On the whole, however, the data are not accurate enough for detailed consideration along these lines.

As another example of the way in which the trace elements distributed themselves among the primary precipitate minerals, the case of the ferrogabbro 1907

* The modal values are too high to represent the amount of the primary precipitate because the outer zones formed from the interstitial liquid are included but the limited accuracy of our data does not justify correcting for this factor.

will be considered (Table 3). In this rock the primary minerals include ilmenite and magnetite in addition to plagioclase, pyroxene and olivine. The analyses show the same general features of trace element distribution as the hypersthene olivine gabbro but the actual figures differ much. At this late stage in the differentiation, the composition of the liquid from which the crystals formed had become impoverished in certain constituents, *e.g.* chromium and nickel, and these are not therefore available to enter the crystal phases in appreciable amount. Only olivine, the mineral which is normally richest in nickel, contains this element above the sensitivity. Other differences from the minerals of the hypersthene olivine gabbro 4077 are no doubt due to changes in the ease or difficulty of the various replacements in the lower temperature solid solution minerals present in this rock. The chief features of the trace element distribution in the minerals of the hortonolite ferrogabbro and comparisons with the case of the hypersthene olivine gabbro 4077 may be listed as follows:

Gallium, besides occurring in the plagioclase enters the magnetite in significant amounts.

Vanadium is still present in the pyroxene but is again not present in the olivine. It is also abundantly present in the magnetite (300 ppm) and is present but in much smaller amount (20 ppm) in the ilmenite. It may be in the plagioclase to the extent of 10 ppm.

Molybdenum is again present in the olivine but not in the pyroxene and there are only small amounts present in the iron ores.

Cobalt is present in pyroxene, olivine, ilmenite and magnetite, in amounts roughly proportional to the ferrous iron content of these minerals.

Copper is present, again about equally, in all the minerals but at ten times the previous level.

Scandium, as in the previous case, is present above the sensitivity only in the pyroxene, but at this stage it is present at five times the previous amount.

Zirconium is again present in the pyroxene and at about the same amount as in the previous case.

Strontium is present abundantly in the plagioclase as in the previous case.

Barium is present in the plagioclase in increased amount.

In comparing the trace elements in the rock with the summation of the amounts in the minerals of the primary precipitate, the same general principles apply as for the hypersthene-olivine gabbro and on the whole the summations are as close as would be expected.

The primary precipitate minerals so far considered are well developed, unzoned crystals which are believed to have formed slowly from a liquid effectively stirred by convection currents. These conditions which permitted optimum regularity of crystal growth, should also have allowed the maximum selective action to be exerted by the growing crystals on the trace elements distributed through the liquid. For comparison with crystals formed under these conditions we have separated the plagioclase, pyroxene and olivine from the chilled marginal gabbro 1724 (see p. 138). In this rock the crystals were formed, presumably without significant stirring of the magma, by diffusion of materials from the magma which immediately surrounded the growing crystals. The plagioclase crystals can be seen to be zoned and no doubt the crystals of pyroxene and olivine are also. Although formed under different conditions it is of interest to find that these minerals have essentially the same trace element composition (Table 4) as the primary precipitate minerals separated from the layered rock 4077 (Table 2). It had been anticipated that, during the more rapid crystallization without stirring, which gave rise to the chilled marginal

Table 3—The trace elements of the hortonolite ferrogabbro, 1907 and of its individual minerals

(Figures in parts per million)

	r	s	*Plag. An 56 (47% in rock)*	*Pyrox. En23 (22% in rock)*	*Oliv. Fo41 (19% in rock)*	*Ilm. (4% in rock)*	*Mag. (8% in rock)*	*Rock*	*Σ minerals*	*Amounts in 47% Plag.*	*Amounts in 22% Pyrox.*	*Amounts in 19% Oliv.*	*Amounts in 4% Ilm.*	*Amounts in 8% Mag.*	*Estimated Composition of magma*
P	0·34		—	—	—	—	—	(350)	—	—	—	—	—	—	(5000)
Ga	0·62	1	50	5	5	3	30	20	28	24	1	1	—	2	22
Cr	0·64	1	*	*	*	*	*	*	—	—	—	—	—	—	1
V	0·65	5	10	30	*	20	300	15	37	5	7		1	24	3
Mo	0·68	1	*	*	10	3	3	*	2	—	—	2	—	—	*
Li	0·78	1	3	*	5	*	*	3	2	1	—	1	—	—	5
Ni	0·78	2	*	*	10	*	*	*	2	—	—	2	—	—	1
Co	0·82	2	*	40	100	70	60	40	37	—	9	20	3	5	18
Cu		10	250	300	300	300	500	400	294	120	65	57	12	40	410
Sc	0·83	10	*	150	*	*	*	10	33	—	33	—	—	—	9
Zr	0·89	10	*	50	*	*	*	20	11	—	11	—	—	—	140
Mn	0·91	10	50	(1700)	(3200)	—	—	(1300)	1000	24	(370)	(610)	—	—	(1750)
Y	1·06	30	*	*	*	*	*	*	—	—	—	—	—	—	65
La	1·22	30	*	*	*	*	*	*	—	—	—	—	—	—	32
Sr	1·27	10	3000	80	100	*	*	700	1440	1400	18	20	—	—	445
Ba	1·43	5	200	10	10	*	*	50	99	95	2	2	—	—	200
Rb	1·49	20	*	*	*	*	*	*	—	—	—	—	—	—	15

An asterisk indicates that the element is present in amounts less than the sensitivity.
The percentages given for the minerals in the rocks are weight percentages.
r is the ionic radius in Ångstrom units at valency shown.
s is sensitivity with spectrographic method employed.

gabbro, the selective action of the growing crystals on the trace constituents of the magma would have been much less effective but within the limits of accuracy of our data this effect is not shown.

Table 4—The trace elements of the chilled marginal gabbro, 1724 and of its individual minerals
(Figures in parts per million)

	r	s	*Plag.* (50% *in rock*)	*Pyrox.* (28.5% *in rock*)	*Olivine* (19.5% *in rock*)	*Rock*	Σ *minerals*	*Amounts in* 50% *plag.*	*Amounts in* 28.5% *pyrox.*	*Amounts in* 19.5% *Olivine*	
P			—	—	—	(**260**)	—	—	—	—	P
Ga	0·62	1	50	5	*	**15**	**26**	25	1·5	—	Ga
Cr	0·64	1	*	600	*	**200**	**170**	—	170	—	Cr
V	0·65	5	*	400	*	**150**	**115**	—	115	—	V
Mo	0·68	1	*	3	10	*	**3**	—	1	2	Mo
Li	0·78	1	*	5	3	**2**	**2**	—	1·5	·5	Li
Ni	0·78	2	2	200	250	**150**	**107**	1	57	49	Ni
Co	0·82	2	3	70	150	**50**	**50**	1	20	29	Co
Cu		10	40	40	100	**150**	**51**	20	11	20	Cu
Sc	0·83	10	*	30	*	*	**8**	—	8·5	—	Sc
Zr	0·89	1	*	50	*	**30**	**14**	—	14	—	Zr
Mn	0·91	1	100	—	—	(**690**)	**50**	50	—	—	Mn
Y	1·06	30	*	*	*	*	—	—	—	—	Y
La	1·22	30	*	*	*	*	—	—	—	—	La
Sr	1·27	10	1000	20	10	**300**	**500**	500	5·5	2	Sr
Ba	1.43	5	20	*	*	**40**	**10**	10	—	—	Ba
Rb	1·49	20	*	*	*	*	—	—	—	—	Rb

An asterisk indicates that the element is present in amounts less than the sensitivity.
The percentages given for the amount of the minerals in the rock are weight percentages.
r is the ionic radius in Angstrom units at the valency shown.
s is the sensitivity with the spectrographic method employed.

VI. Distribution of Trace Elements in the Chief Mineral Series

In this section the trace element composition of the primary precipitate minerals is considered and especially the changes taking place as the solid-solution mineral series change in composition during fractionation. The composition of the three primary precipitate minerals of the hypersthene olivine gabbro 4077, discussed in the previous section, will be taken as a basis for comparison and with them the earlier, higher temperature, and the later, lower temperature solid-solutions will be compared. The trace elements in certain minerals of the hybrid rocks are also considered (Section 6) although nothing like a full study of them has yet been made. The effect of various types of impurity on the data for the minerals is considered in the appendix.

1. Plagioclase Series

Of the plagioclases analysed spectrographically, that from the hypersthene olivine gabbro 4077 was obtained in fragments of high purity except for certain minute inclusions (types II, III and sometimes IV, as defined in the appendix) and the others were similar except that they contained rather more inclusions. Chemical analyses for the major constituents of three of the plagioclases have been kindly undertaken by Dr. W. A. Deer. The whole of the spectrographic and chemical data available are assembled in the first section of Table B. The figures for the spectrographic data in columns I and II of this table are the mean of two determinations.

There are significant amounts of ferric iron in the three chemically analysed plagioclases. There is also a smaller, but probably significant amount of ferrous iron and magnesium. A calculation of the number of metal atoms associated with eight oxygen atoms has been made (Table 5) and when the ferric iron is considered as a replacement of Al, this group approaches closely to the ideal value of four. In these plagioclases it seems clear that there is a significant replacement of Al by Fe^{+++}.

Table 5—Formulae of three analysed plagioclases

	Plagioclase 4077			*Plagioclase* 2580			*Plagioclase* 4145		
Normative Comp. (Wt.%)	$An_{55}\ Ab_{43}\ Or_2$			$An_{41}\ Ab_{56}\ Or_3$			$An_{38}\ Ab_{60}\ Or_2$		
Comp. from μ (Wt.%)	An57			An41			An38		
	Mol. Props.	*Atom ratios to 8 oxygens*		*Mol Props.*	*Atom ratios to 8 oxygens*		*Mol Props.*	*Atom ratios to 8 oxygens*	
SiO_2	0·906	**2·470**	**4·002**	0·967	**2·611**	**3·993**	·979	**2·651**	**3·998**
Al_2O_3	0·278	**1·516**		0·253	**1·366**		·245	**1·320**	
Fe_2O_3	0·003	**0·016**		0·003	**0·016**		·005	**0·027**	
FeO	0·003	**0·008**	**1·017**	0·004	**0·011**	**0·999**	·006	**0·019**	**0·987**
MgO	0·0005	**0·001**		0·001	**0·002**		·0005	**0·001**	
Na_2O	0·080	**0·436**		0·104	**0·562**		·109	**0·587**	
CaO	0·200	**0·545**		0·143	**0·386**		·127	**0·345**	
SrO	0·002	**0·005**		0·004	**0·011**		·003	**0·008**	
K_2O	0·004	**0·022**		0·005	**0·027**		·005	**0·027**	

General formula taking into consideration data from this table and Table B (Ca, Na, K, Sr, Fe··, Ba), (Si, Al, Fe···, Ga)$_4O_8$.

The possibility of Fe^{+++} replacing Al in orthoclase was shown by the early work of Hautefeulle and Perrey (1888) who reported synthesizing an iron orthoclase. Later Lacroix (1922, pp. 560–561) described a yellow orthoclase from Madagascar containing 12% of ferric iron orthoclase and Faust (1936) has investigated the conditions for the synthesis of iron orthoclase and discussed its occurrence in nature. The possibility of the presence of Fe^{+++} in the albite-anorthite series has been indicated by the syntheses of Day and Allen (1905, p. 38) and has since been discussed by Schiebold (1931) and Faust (1936). No synthesis of ferric anorthite or albite has been made nor have iron-rich varieties of natural plagioclases been described. From the evidence of the analyses given here and of other good quality analyses (*cf* Alling, 1921) it appears that limited replacement of Al by Fe^{+++} occurs in the plagioclases under natural conditions, the extent of the replacement reaching about 0·5% Fe_2O_3. It is of interest to note that the amount of ferric iron in plagioclase from the gabbro 4077 is greater than the amount in the pyroxene from the same rock.

The amount of ferrous iron in the plagioclases (0·2 to 0·4%) is also considerable and it does not seem likely that it can be due to accidental impurities or injected inclusions as defined in the appendix. The Fe^{++} and also the Mg is probably present in the felspar replacing Na and these elements are so shown in Table 5. Reheating of plagioclase under certain conditions of thermal metamorphism produces a clouding of the plagioclase felspars which has been described by MacGregor (1931). The clouding is due to the development of minute foreign crystals which MacGregor suggests are usually iron ore. This clouding is presumably an ex-solution effect due

to the separation of some of the iron originally present in the plagioclase structure. In the Skaergaard analyses it is interesting to note that the amounts of Fe_2O_3 and FeO shown by the analyses are roughly in the proportion required to form magnetite, but this is perhaps a coincidence.

The amount of K_2O present in these plagioclases is small; indeed it is only about the same as the amount of ferric iron. During fractionation the change in the amount of K_2O is also small, and this is in harmony with data summarized graphically by Alling (1921)*.

In the course of fractionation, while the plagioclase changes in general composition from An65 to An30, the chief features of the trace element composition are as follows:

Gallium remains steady at about 50 ppm until the latest fraction when there is an abrupt rise to 200.

Chromium is apparently present to a small extent in the earliest plagioclases, when this element is relatively abundant in the magma.

Vanadium seems to be present in small amount from about 5 to 20 ppm. If Fe^{+++} enters the crystal, then small amounts of V and Cr might be expected.

Lithium is shown by our results to be present in amounts just above the sensitivity and there is no particular change during fractionation. It may not be truly present in the crystal structure; the small amounts found may be the result of additions during slight late-stage alteration of the plagioclases.

Copper is present in small amounts, about 10 ppm, in the earliest plagioclase and there is marked enrichment in successive plagioclase fractions until, in the ferrohortonolite ferrogabbro stage it amounts to 700 ppm. After this, in the latest analysed plagioclase, the amount of copper falls abruptly to 10 ppm. The steady increase and abrupt fall in the amount of copper shown by the successive plagioclases is also shown by the pyroxenes and olivines. The sudden fall in copper content of the silicate minerals does not correspond to a fall in the copper content of the rocks, but it has been found that, at the stage when the amount of copper in the silicate minerals becomes much reduced, sulphides become abundant in the rocks. If the sulphide formed a separate immiscible liquid phase, the bulk of the copper would pass into it leaving the silicate liquid impoverished, and this would necessitate the chief minerals formed from the silicate liquid having a reduced copper content. Until the abrupt appearance of rather abundant sulphides in the later ferrogabbros was discovered, the behaviour of the copper had seemed inexplicable (*cf* Wager and Mitchell, 1950, p. 147). The possibility of an immiscible sulphide phase and its relation to the copper contents of the rocks and minerals is discussed more fully later (p. 182 and pp. 190–192).

Manganese is present in the felspars in appreciable amount. The data indicate higher values in the later stages when the amount of Mn in the liquid is considerably higher.

Strontium is about 1000 ppm in the earlier plagioclases; in later ones the amount rises, reaching 5000 ppm, but in the latest the amount falls to 2000 ppm. Since the strontium content is somewhat high for our spectrographic method of determination there may be rather considerable errors in the absolute amounts, although the trend of variation indicated by our figures is no doubt correct. A chemical determination of the strontium in the plagioclase An37 by W. A. Deer gave 2500 ppm of strontium while the spectrographic value is 5000 ppm. Perhaps the true figure lies between these values.

Barium is low in the early plagioclases and rises steadily until, in the latest, it is ten times as abundant. In view of the relatively slight increase in potassium in this series it is interesting that barium should increase so emphatically.

An approximation to the composition of the liquids from which the successive minerals separated may be obtained from the graphs (Figure 7) which are based on the estimated contents of certain specific residual magmas (pp. 174–177 and Table 10). Using these data for the composition of the liquids and the data for composition

* In connection with the chemical analyses it should be noted that when they are recalculated to show the proportions of orthoclase, albite and anorthite units then the composition corresponds well with that previously estimated from refractive index determinations (see top of Table 5).

of the plagioclase crystals from Table B, it is possible to calculate approximate distribution factors between crystal and liquid for the various elements at the different stages. The factor used is the amount by weight of the element in the crystal divided by the amount by weight in the liquid; thus, when this factor is greater than unity, the crystal phase shows enrichment relative to the liquid, and when less than unity, the crystal shows impoverishment. The distribution factor gives an estimate of the tendency for a particular element to enter the crystal structure. Usually this tendency has been expressed qualitatively by saying that the element enters the crystal structure "readily" or "to some extent", *etc.* In making such statements, however, the change in the amount of an element in the various crystals has often been taken as a sufficient indication of the tendency of the element to enter the crystal phase, without reference to the concentration in the liquid from which the crystal is separating. The distribution factor takes into consideration the concentration in the liquid and is a better measure of the tendency of trace elements to enter the crystal structure.

Considerable irregular variation in the distribution factors is to be expected because of the limited accuracy of the spectrographic data, and because of difficulties in estimating the composition of the successive magmas. On the whole the errors in estimating the magma composition is not likely to produce significant errors in the value of the distribution factors in the A to C stages, but errors probably become more serious in the later stages, especially when the amounts of the elements present are very low. The distribution factors for the plagioclases are given in Table 6, and we shall give brief comments on them here, reserving until later the general discussion of their significance.

The distribution factors for those elements which probably replace aluminium within the oxygen tetrahedra may first be considered:

For *Gallium*, the distribution factor is about 2, like that for Al, until the latest plagioclase when it rises to 7. A rise in the amount of gallium relative to aluminium in late fractions has been shown by Goldschmidt, an enrichment which, he points out, is in harmony with the greater ionic size of Ga. In our series there is apparently a rise in the distribution factor towards the end of the fractionation implying greater ease of entry into the latest plagioclase.

Ferric Iron has a distribution factor of about 0·1 while that for Al is about 2, *i.e.* Fe^{+++} has a much smaller tendency to enter than Al. Fe^{+++} is present in chemically analysable amount only because there is a considerable amount of ferric iron in the liquid.

Chromium and *Vanadium*, have some tendency to replace Al, although this tendency is even less than for Fe^{+++}. The distribution factors for V are shown to increase rapidly in the successively lower temperature plagioclases, except for the latest. This increase, we believe, is a real tendency, the latest figure being unsatisfactory because of a poor estimate of the content of V in the magma.

The other elements, when truly present in the structure and not there by contamination, are probably replacing the Na and Ca ions outside the SiO_4 network and these may now be considered:

Magnesium, *Ferrous Iron* and *Nickel* all have very low distribution factors, round about 0·01. Only Ni appears above the sensitivity for the first plagioclases and it might be thought of as an accidental impurity, except that its distribution factor is of the same order as for Fe^{++} and Mg, with which it should show similarity. Moreover this felspar was separated with great precautions and analysed twice, so that we believe the Ni is really present in the structure. Probably nickel and cobalt have a similar tendency to enter plagioclase as Fe^{++} and Mg, but as they are present in the magma in such small traces, the amount actually entering is below the sensitivity except for Ni in the first plagioclase.

The distribution factors for *Copper* definitely indicate a greater readiness to enter than is the case for Fe^{++} and Mg, and the values for the distribution factor increase in the later plagioclases. The figure given for copper in the latest felspar from the fayalite ferrogabbro is placed within square brackets as the value given for copper in the magma would be wrong if there were a development of an immiscible sulphide phase.

The distribution factors for *Manganese* show considerable variation, but on the whole the figures are higher than for Fe^{++} and it appears that the larger Mn^{++} is able to enter more easily than Fe^{++}.

Strontium shows high distribution factors varying from 2 to 6 and possibly more (the figure of 12 may be high due to some uncertainty in estimating high Sr contents). These values are significantly higher than those for calcium. Since the Sr ion is considerably larger than the Ca, this behaviour is unexpected but is in harmony with the higher melting point of strontium anorthite.

Potassium shows distribution factors which decline with differentiation from 1·3 to 0·8. It is clear that the data on compositions summarized by Alling (1921, Figure 19) are in harmony with the change in distribution factors for potassium here given. Thus he shows by a graphical presentation of the available analyses that potassium is less abundant in the soda-plagioclases than it is in the middle range and, since the more acid plagioclases have normally separated from liquids considerably richer in potassium than those from which basic plagioclase separated, the potassium distribution factor in the acid plagioclases must be on the average lower than for the intermediate plagioclases.

Barium shows distribution factors fairly similar to those for K. They are markedly less than those of Sr. It is interesting to note that the distribution factors for barium remain fairly steady over the whole range of these plagioclases, while the actual amount of Ba in the plagioclases increases ten times. This is a clear reminder that we must not speak of a stronger tendency for barium to enter the later plagioclases simply on the basis of an increase in the barium content without considering the changes in composition of the magma. A low temperature albite which is not a mineral of the layered series, but from a miarolitic cavity, is of interest in showing a much reduced amount of Ba (see below p. 48). This may perhaps be taken as evidence that the distribution factor for Ba also falls in the late stages but this cannot be stated definitely because the composition of the liquid from which it separated is unknown.

The results presented in this section on the non-essential or trace element constituents of the plagioclases which range from An56–An37 may be summarized as follows :

Fe^{+++} is a significant constituent apparently replacing Al in tetrahedral co-ordination. The evidence, though less definite, also suggests that chromium and vanadium replace Al in this way.

Ga also replaces Al and, judged by the distribution factors, the tendency is apparently constant up to the latest plagioclase when there is a marked increase.

Fe^{++}, Mg, Cu and Mn are regarded as significant constituents replacing Na or Ca, but only in minute amounts.

The ease with which K enters, compared with its concentration in the magma, is slightly greater in labradorite than in andesine and oligoclase. In two of the plagioclases K_2O is actually less abundant by weight than Fe_2O_3.

Sr is an important constituent throughout, reaching a maximum in the andesine. It has a stronger tendency to enter than has Ca.

Ba enters less readily than Sr. It is ten times more abundant in the latest plagioclase than in the earliest, but the amount in the magma has increased in about the same proportion and the tendency to enter, as judged by the distribution factor, is fairly constant over the range shown by the layered series plagioclases.

Li is present in the samples analysed. Whether it be part of the crystal structure or introduced later with slight hydrothermal alteration is not clear.

2. *Clino-Pyroxene Series*

From the varied major element composition shown by pyroxenes it is clear that the structure is tolerant of a variety of ions which mainly occupy the Ca and Mg positions of diopside. The chief replacements shown by ordinary chemical analysis are: Fe^{++} and Mn^{++} for Mg^{++}; (Na^{+}, Fe^{+++}) for (Ca^{++}, Mg^{++}) and, in this latter type of replacement, Na may apparently be replaced to some extent by K^{+}, and Fe^{+++} by Al^{+++} and Ti^{++++}; in addition Al^{+++} can replace Si in the silicon oxygen tetrahedra and there is some evidence that Ti may also do the same (Barth, 1931). If this is so, and in view of the replacement of Al^{+++} by Fe^{+++} in the plagioclases discussed above, it would seem likely that some replacement of Si by Fe^{+++} might also occur in the pyroxenes. The formula for pyroxene from the fayalite ferrogabbro 1881 has been given in a previous paper as (Mg, Fe^{++}, Fe^{+++}, Ca, Mn, Na, K, Ti, Al)$_2$ (Si, Al)$_2O_6$ (Deer and Wager, 1938, p. 16).

With high tolerance for considerable quantities of ions of varying character it would be expected that a great variety of trace elements would enter the structure. This has been shown by analyses made by Goldschmidt and his collaborators and is confirmed by the present analyses. However, the order of entry of trace elements into the pyroxene structure during fractional crystallization is a matter about which less is known and prediction at present from crystal-chemical concepts would be hazardous. The data here presented may be taken as giving a good indication of the way in which trace elements enter the diopside-hedenbergite pyroxene series under conditions of strong fractionation.

In the pyroxenes of the layered series which range from Wo_{41} En_{42} Fs_{17} to Wo_{30} $En_{1\cdot5}$ $Fs_{68\cdot5}$ there is so strong a variation in the trace elements that there are few common features between the extremes of the series (Table B). The first two analyses (columns A, B) may be considered ordinary gabbroic pyroxenes and both contain Cr, V, Ni, Co, Cu, Sc, Zr and Mn as significant trace constituents. The amount of Sr is surprisingly low. On the other hand, the latest, iron-rich pyroxene has less than 2 ppm of Cr, V or Ni, the significant trace constituents among those we have sought being Co, Li, Sc, Mn, Y and La.

The variation in amounts of the different elements in the pyroxene series together with comparisons between the composition of the crystals and of the liquid from which they formed will now be considered element by element in order of increasing ionic size:

Chromium is present in the earliest pyroxene to the extent of 3000 ppm. The chromium content of the liquid from which this pyroxene separated was only 170 ppm and thus the concentration of chromium in the pyroxene is eighteen times that in the liquid (for the distribution factors see Table 6). In the somewhat more iron-rich pyroxene from 4077 there is still 350 ppm chromium and this is estimated to represent a concentration in the crystal nine times that in the liquid. The more iron-rich pyroxene from the middle gabbro (column C) has less than 1 ppm chromium and so have the later pyroxenes.

Vanadium enters the early pyroxene in moderate amount and persists to the middle of the solid solution series. The distribution factor is about 2 over the range when the value is most certain and this is not so high as for chromium (distribution factor about 10). In the later pyroxenes of the fractional crystallization series V is below the sensitivity, *i.e.* < 5 ppm.

Lithium is present only in amounts near the sensitivity value except in the latest pyroxene when the amount rises abruptly to 50 ppm. This is in harmony with the work of Strock (1936) who has shown that in the late stages of crystal fractionation Li has a strong tendency to occupy sites normally filled by Mg and Fe. This has been accounted for by the lower charge on the Li ion, which is unable to gain access to the pyroxene structure until concentrations of the other competing elements are reduced.

Nickel occurs in the early pyroxenes and persists to the middle of the series but is absent in the later. It is clearly replacing Mg but the amount of Ni becomes reduced below 1 ppm in the hortonolite ferrogabbro (column D) while there is still 10% Mg present. The distribution factor found for Ni in the early pyroxenes is 1·2 and this rises to 5 in Middle Gabbro pyroxenes while the distribution factor for Mg is still only 2·5. It seems probable that the Ni content of the early pyroxene is under-estimated and that the distribution factor should be about 3. The relation between Ni and Mg, expressed by the ratio Ni/Mg $\times 1000$ is 2·1 for the earliest pyroxene and changes successively with fractionation to 1·6, 0·7, $< 0{\cdot}03$. There is thus a violent change in the relative amounts of Ni and Mg. The sizes of the Mg and Ni ions in six-fold coordination are essentially the same, and Goldschmidt (*cf* 1944, p. 4) has indicated that the early concentration of Ni relative to Mg is due to polarization of the Ni ion.

The *Cobalt* content of the pyroxenes remains steady at about 50 ppm over most of the range, only falling slightly to 15 ppm in the most iron-rich pyroxenes. Relative to Fe^{++} the amount of Co falls throughout the series. The distribution factor ranges from 1·0 to 2·5 much as for Fe^{++}. The ratio Co/Ni varies significantly with fractionation being about 0·3 in the gabbroic pyroxenes, 1 in the Middle Gabbros and > 40 in the later iron-rich pyroxenes.

From its general ionic size *Copper* might be expected to replace Fe in ferro-magnesian minerals and this it seems to do, reaching 1000 ppm in the pyroxenes of the ferrohortonolite ferrogabbro stage. The abrupt fall in amount of copper in the pyroxene of the fayalite ferrogabbros is considered due to the formation of an immiscible sulphide liquid (see under plagioclase p. 147). There is a steady rise in the distribution factor until this stage is reached. If there be immiscibility in the liquid phase the value given for the distribution factor of the latest pyroxene will not be the true factor.

Scandium is present in the pyroxenes and shows little significant change throughout the series. This element enters pyroxene much more readily than it enters plagioclase, olivine or the iron ores. The distribution factor lies between 3 and 20. Among igneous minerals the usual habitat of scandium is the pyroxene of basic rocks as was shown by Goldschmidt and Peters (1931) (see also Goldschmidt, 1934, pp. 395–397).

Zirconium enters the early and middle pyroxenes and is below the sensitivity in the two later pyroxenes analysed. It is likely that Zr replaces Ca in the pyroxene structure.

Manganese in the pyroxene has usually been determined by chemical analyses, as it is a fairly abundant constituent. It has apparently a distribution factor similar to, or rather higher than Fe^{++} and Ca. On the basis of size it would be expected to enter less readily if replacing Fe and more readily if replacing Ca. Perhaps the Mn occupies both the Mg and Ca positions of diopside equally readily. In this connection it is interesting to note that the amounts of Mn in the olivines, where there is no Ca to replace, is much the same as the amounts in the pyroxenes.

Yttrium and *Lanthanum* are clearly concentrated in the later pyroxenes and must be thought of as replacing Ca. The figure of 200 ppm of La in the earliest pyroxene is anomalous.

Strontium is relatively low in amount in all the pyroxenes although the data suggest a slight increase in Sr in the later fractions. In the early and middle parts of the differentiation series there is 20 to 100 times as much strontium in the plagioclases as in the pyroxenes. It is most striking how the strontium can enter abundantly the early plagioclases, presumably replacing Ca, whereas it cannot replace the Ca in the early pyroxenes except in very small amounts.

Noll (1934) showed a similar contrast in Sr values for plagioclases and pyroxenes separated from basic rocks and similar results are shown by Nockolds and Mitchell (1948) for plagioclase and pyroxene separated from a pyroxene-mica-diorite. That there would be this contrast between the entry of Sr into diopsidic pyroxenes and into plagioclase was foreshadowed from the experimental work of Eskola (1922), see below p. 194. Comparisons between the ease of entry of Sr and K, as indicated by the distribution factors is interesting; the figures seem to be reasonably satisfactory and to show that K enters more readily than Sr.

There is only a small amount of *Barium* in the pyroxenes but some of that shown by our analyses may well be due to contamination as roughly similar amounts are present in the olivines.

Rubidium is reported in the latest pyroxene. It is possible that this is due to slight alteration by residual watery solutions in which the Rb had become concentrated.

In ordinary pyroxenes of basic rocks the presence of chromium and nickel has long been known. Goldschmidt's work showed that scandium entered readily. Our data show vanadium and zirconium as further trace elements characteristic of ordinary basic pyroxenes. The iron-rich pyroxenes produced by strong fractionation seem to accommodate Li, Y and La with some Co and Sc, but Cr, V, Ni and Zr are virtually eliminated. These generalizations being based on a diopside-hedenbergite series of pyroxenes do not give much indication of the trace element constituents of other types of pyroxene with a different paragenesis. The actual trace element composition of alkali pyroxenes and of pyroxenes from granites requires direct investigation and we hope to attempt the latter on pyroxenes from Skye granites.

Unfortunately the Skaergaard intrusion does not provide any satisfactory data on the ortho-pyroxenes. They are only present in subordinate amounts and never become primary precipitate minerals so that we have not attempted their separation for analysis. The early clino-pyroxenes which we have analysed spectrographically probably contain a few per cent of ortho-pyroxene, an amount considerably less than that in the material analysed chemically for the major elements as hand picking no doubt largely eliminates it (*cf* 1939, pp. 76–83). In dealing with the spectrographic data it was not thought necessary to make any allowance for the small amounts of ortho-pyroxene. No doubt ortho-pyroxenes have a somewhat different trace element make-up from that of the clino-pyroxenes and Nockolds and Mitchell (1948, p. 555) have given some evidence for this.

3. Olivine Series

Examination of the analyses of the separated olivines (Table B) shows at once that the olivine structure is much less able to accommodate a variety of ions than the pyroxene. In the gabbroic olivines Mo, Li, Ni, Co, Cu and Mn are present. In the successively more iron-rich olivines produced by fractionation, Ni declines rapidly until there is < 2 ppm, Co slowly, while Li, Cu, Mn and Mo increase markedly.

The various trace elements shown by the olivine analyses will be considered as previously in the order of increasing ionic size:

Chromium is shown as present to the extent of 20 ppm in the first olivine. Impurity of one part in 150 of pyroxene would account for this and it is not considered that our data give a definite answer to the question of whether or not a small amount of Cr may be in the olivine structure. The chemical analyses show some Fe_2O_3 in the olivines but it is also not clear whether this is truly in the crystal structure. A common feature of the Skaergaard olivines and, indeed, of many olivines of plutonic rocks, is the existence of orientated opaque inclusions usually considered to be iron ore formed by an ex-solution process. These inclusions may account for the Fe_2O_3 in the analyses and minute amounts of Cr_2O_3 may also be present with the Fe_2O_3. Assuming that the inclusions are due to ex-solution, then the material would have been in the structure at high temperatures when there was greater size tolerance for ions. It might be anticipated that Cr^{+++} would be present with the Fe^{+++} in the SiO_4 tetrahedra as apparently in felspar, but if this were so a greater entry of Al would have been expected. Two especially complete analyses of olivines were recently made by Goldich (1943, p. 177). They are of forsterite-rich olivines and they also contain ferric iron and small amounts of Cr_2O_3. At an early stage of this work we separated an olivine from the gabbro picrite which should have provided an example of an early fraction of olivine. Unfortunately the sample could not be obtained free from occasional opaque grains and dusty opaque material produced along cracks by decomposition. This olivine had the same trace element composition as that from 1846 (column A) except that 1000 ppm Cr were found. Re-examination of thin sections of the rock shows minute clusters of an opaque mineral regarded as chrome spinel. The minute crystals have the peculiar synneusis distribution as described by Vogt (1921, pp. 321–322). It appears that a chrome spinel was directly precipitated from the magma at an early stage.

a point considered more fully later (p. 183). The presence of chrome spinel as an impurity in the olivine is regarded as the cause of the high chromium content of the olivine sample from the gabbro picrite. For this reason the analysis is not given here although it was quoted in part in an earlier note (Wager and Mitchell, 1945).

Molybdenum seems to be a real, but unexpected trace element of the olivines. Inspection of the range of analyses shows that it cannot be present as a result of contamination by any of the analysed minerals, and it is not present equally in all the minerals so that it is unlikely to be due to the so-called injected inclusions (appendix p. 203), though it might perhaps be due to residual solutions altering the olivine along cleavage cracks. We are, however, of the opinion that it is really present in the crystal structure.

The amount of *Lithium* is comparable with that in the pyroxenes.

Nickel enters early olivines in abundance, as is well known (*cf* Vogt, 1923). Judging by the distribution factors, nickel enters early olivines with greater readiness than Mg, the distribution factor for the early gabbroic olivines being 10 for Ni while for Mg it is about 5 (Table 6). The amount of Ni in the olivine series is reduced below 1 ppm at the stage when the amount of Mg is still 8%; there is thus a marked change in the Ni/Mg ratio.

Cobalt, as in the pyroxenes, is not strongly concentrated, and it shows a decline in amount in the later olivines. The distribution factor remains between 3 and 5 over a long range and is on the whole greater than that for Fe^{++}.

Comparison of the nickel and cobalt contents of pyroxenes and olivines from the same rock shows that the amounts of these elements in the olivines are about double those in the pyroxenes. This is no doubt because the amount of the trace element entering is roughly proportionate to the amount of Mg+Fe in olivine which is approximately twice that in pyroxene.

Copper is present in the olivines and rises steadily with fractionation until it reaches about 400 ppm in the ferrohortonolite ferrogabbros. The amount then falls abruptly as with the plagioclases and pyroxenes and for the same reason (see p. 147).

Manganese rises in the successively more iron-rich olivines. The distribution factor is much the same as for Fe^{++} and on the whole it increases slightly with differentiation.

The amounts of *Strontium* and *Barium* are comparable with those in the pyroxenes and it would have been expected that replacement of the calcium of the pyroxene by these elements would have resulted in a greater concentration in the pyroxenes than the olivines. The similarity in amounts of Sr and Ba in pyroxenes and olivines suggests the possibility that there has been contamination by plagioclase material and that the figures are not significant.

It is worthy of comment that the amounts of *Vanadium*, *Scandium* and *Zirconium* are all below the sensitivities while these elements are present in significant amounts in the pyroxenes.

4. Ilmenite and Magnetite

The ilmenite and magnetite samples were separated from the crushed rocks by a simple magnetic method. From the earlier investigation (1939, p. 230) it was known that the two ores occurred in essentially separate crystal masses but considerable contamination in the separated samples was anticipated as no visual test of homogeneity by transmitted light could be applied while hand picking.

Since making the separations the nature of the ore minerals has been examined in polished specimens, with the help of Dr. H. Neumann of Oslo University. The two lowest rocks from which the iron ores were separated (2308 and 2362) show that ilmenite and magnetite are present in well defined, big, individual crystals and no ex-solution lamellae of one in the other are visible (Figure 4). It is interesting that the ilmenite and magnetite grains have a tendency to cluster together in bunches free from silicate minerals suggesting that these minerals even at the high temperatures of precipitation had a feeble magnetism. For these two rocks the magnetite and ilmenite samples which have been analysed are no doubt reasonably free from contamination one by the other and indeed consideration of the amount of chromium

Table 6—Estimated distribution factors (crystal/liquid)

The ratios have been calculated from the composition of the minerals given in Table B, and the estimated composition of the magma taken from the graphs of Figure 7 and given on the left hand side of the table. Figures which are believed, for various reasons, to be misleading are put in brackets.

		Estimated composition of the magmas corresponding to the various rock stages					
	r	*A*	*B*	*C*	*D*	*E*	*F*
		Weight per cent of oxide					
SiO_2	0·39	47·92	46·5	45·5	45·0	47·0	49·0
Al_2O_3	0·57	18·87	15·0	13·5	12·5	12·0	13·0
TiO_2	0·64	1·40	2·4	2·4	2·3	2·0	1·6
Fe_2O_3	0·67	1·18	3·1	3·5	3·9	4·9	5·5
MgO	0·78	7·82	5·5	5·0	3·0	1·2	1·0
FeO	0·83	8·65	14·4	16·5	19·1	20·1	17·0
Na_2O	0·98	2·44	3·3	3·1	3·2	3·1	3·0
CaO	1·06	10·46	9·5	8·5	8·0	7·5	7·0
K_2O	1·33	0·19	0·35	0·4	0·5	0·8	1·0
		Parts per million of the element					
Ga	0·62	17	20	20	22	23	28
Cr	0·64	170	50	1·0	1·0	1·5	4·0
V	0·65	140	170	45	3	1·2	3
Mo							
Li	0·78	2	4	5	5	8	17
Ni	0·78	170	35	10	1·0	1·5	4·0
Co	0·82	53	30	25	18	13	6
Cu		130	245	310	410	440	[330]
Sc	0·83	12	13	10	9	8	2
Zr	0·89	50	90	100	140	180	500
Mn	0·91	(800)	(1250)	(1400)	(1750)	(2000)	(1800)
Y	1·06	*	20	35	65	100	250
La	1·22	*	12	17	32	55	160
Sr	1·27	350	495	475	445	430	405
Ba	1·43	43	100	140	200	230	700
Rb	1·49	*	5	10	15	20	40

		Olivine distribution factors					
	r	*A*	*B*	*C*	*D*	*E*	*F*
SiO_2	0·39						
Al_2O_3	0·57	—		—			
TiO_2	0·64	—		—			
Fe_2O_3	0·67	—		—			
MgO	0·78	—	5·5	—	6·0	7·0	(1·0)
FeO	0·83	—	2·2	—	2·5	2·9	3·8
Na_2O	0·98	—		—			
CaO	1·06	—		—			
K_2O	1·33	—		—			
Ga	0·62						
Cr	0·64	(0·12)					
V	0·65						
Mo							
Li	0·78	1·0	0·8		1·0	0·6	1·0
Ni	0·78	12	9·5		10	(<1·5)	
Co	0·82	3·0	4·1		5·5	3·8	3·3
Cu		0·1	0·08		0·7	0·9	[0·06]
Sc	0·83						
Zr	0·89						
Mn	0·91	2·5	1·3		1·8	3·3	4·3
Y	1·06						
La	1·22						
Sr	1·27						
Ba	1·43						
Rb	1·49						

Plagioclase distribution factors

A	*B*	*C*	*D*	*E*	*F*
—		—			—
—	1·9	—	2·0	2·1	—
—		—			—
—	0·16	—	0·1	0·15	—
—	0·004	—	0·013	0·017	—
—	0·013	—	0·017	0·02	—
—	1·5	—	2·0	2·2	—
—	1·2	—	1·0	0·95	—
—	1·1	—	0·9	0·6	—
2·0	2·5	2·5	2·3	1·8	7·0
0·06	< 0·02				
0·07	< 0·03	0·2	3·0	8·0	< 1·5
1·0	1·0	0·5	0·5	0·4	< ·06
0·06	< 0·05				
	< 0·07				
0·1	0·14	0·05	0·6	1·6	[0·03]
	< 1·0				
	< 0·1	< 0·1	< 0·1	< 0·06	0·06
0·13	0·02	0·03	0·03	0·1	0·06
					< 0·1
					< 0·2
3·0	2·0	6·0	6·0	12·0	5·0
2·0	0·6	1·4	1·0	1·7	1·0

Pyroxene distribution factors

A	*B*	*C*	*D*	*E*	*F*
0·19	0·25	0·34	0·24	0·33	0·40
0·6	0·4	0·7	0·3	0·8	1·0
1·6	0·07	0·2	0·38	0·24	0·68
2·0	2·7	2·4	3·5	3·7	(0·47)
1·0	0·9	1·1	0·9	1·2	1·8
0·3	0·13	0·17	0·15	0·17	0·33
1·8	1·8	1·5	2·0	2·3	1·7
0·7	0·4	0·3	0·2	0·25	0·2
0·18	0·15	0·05	0·23	0·4	0·35
18	7	(< 1)			
2·1	1·5	2·2	10·0	< 4	
1·5	0·5	< 0·2	< 0·2	0·3	3
1·2	4·0	5·0	(< 2)	(< 1)	(< 0·5)
1·1	1·7	2·4	2·2	2·3	2·5
< 0·1	0·14	0·3	0·7	2·3	[0·3]
6·0	2·3	5·0	17	10	40
0·6	0·5	0·3	0·35	< 0·05	< 0·02
1·9	1·7	2·0	1·1	1·6	3·3
					2·0
20				3	1·9
0·03	0·02	0·04	0·18	0·07	0·25
0·1	0·07	0·07	0·05	0·02	0·08
				< 1·0	1·0

Ilmenite distribution factors

BC	*C*	*D*	*E*	*F*
0·25	0·15	0·13	0·2	
0·2				
6·0	6·5	6·6	< 5	
7	5	(< 2)		
3	4	4	2·3	
0·3	0·1	0·7	0·9	

Magnetite distribution factors

BC	*C*	*D*	*E*	*F*
1·0	1·5	1·4	2·6	1·1
20				
20	18	20	< 5	< 2
15	10	(< 2)		
3	3·2	3·3	2·3	< 0·3
0·07	0·2	1·2	0·7	[1·2]

in the ilmenite and magnetite from rock 2308 shows that the contamination of the ilmenite by magnetite cannot exceed one part in a hundred. In the ferrogabbros the ilmenite and magnetite still occur in distinct, big crystals but ilmenite ex-solution lamellae* are visible in the magnetite grains (Figure 4). In separating the material for analysis from the hortonolite ferrogabbro, 1907, and from the ferrohortonolite ferrogabbro, 4145, fairly effective separation into ilmenite and into magnetite with ilmenite lamellae has probably been effected. In the case of the magnetite from 4142 the sample is best described as a concentrate of magnetite and not the pure mineral; the polished specimens show that in 4142 the ilmenite and magnetite grains are somewhat more intimately mixed, and the lamellae of ilmenite in the magnetite are abundant. Also pyrrhotite which occurs in this rock, sometimes closely associated with the iron ores, was probably present in the analysed sample.

The iron ores do not form as primary precipitate minerals until a middle stage of the fractionation (Figure 3). The examination of the polished specimens confirms our earlier views (1939, p. 230) that in the early stage of iron ore precipitation, both ilmenite and magnetite separated in approximately equal amounts. The available rock collections are not sufficient to allow us to trace in detail the incoming of ilmenite and magnetite as primary precipitate minerals, but they do show that both these minerals must have become primary precipitate crystals at about the same time—a stage rather later than the hypersthene-olivine gabbro 4077 and yet before the middle gabbro 3662. The iron ores become primary precipitates while the rock is hypersthene-olivine gabbro, and so the earliest, analysed, ilmenite and magnetite belong to the B stage of differentiation (p. 136) although separated from rock 2308 which occurs some 350 m higher in the layered series than rock 4077. With fractionation the amount of ilmenite decreased relative to magnetite (1939, p. 230) and at the same time ilmenite lamellae developed in the magnetite crystals by ex-solution. Unfortunately no chemical analyses of the iron ores have yet been made, but it is intended in the near future to separate the magnetite and ilmenite from these rocks and determine the variation in Al, Ti, Mg, Mn, *etc.*

Iron ores separate from igneous magmas under a wide range of conditions and considerable variation in their trace constituents may be anticipated, while iron ores of still other paragenesis have still wider variations in their trace constituents as shown by Landergren's recent work (1947). The Skaergaard data, however, are of interest because of the steady conditions of fractionation. The chief features shown by the trace constituents are here summarized as for the previous mineral series:

Gallium is a significant trace element of magnetite but not of ilmenite. The amount in the magnetites shows little variation during fractionation. Over the greater part of the range, the quantity present is not greatly different from that in the contemporaneous plagioclase, but in the latest magnetite there is no rise in Ga content as there is in the latest plagioclase.

When ilmenite and magnetite first formed as primary precipitates, the amount of *Chromium* in the magma had already been much reduced by precipitation in early chromite and pyroxene, and this is reflected in the relatively small amount of Cr in the iron ores. Cr occurs in the first magnetite analysed to the extent of 200 ppm and the amount in the ilmenite is only 2 ppm. The trivalent ions Fe^{+++} and Al^{+++} do not enter ilmenite readily, the maximum amount of Fe_2O_3 being 6% according to the authors of the new edition of Dana (1944). This would suggest that the entry of chromium into the

* It is interesting to note that Ramdohr considers that in normal basic rocks unmixing to give ilmenite lamellae takes place unless there is rapid cooling (1940, p. 13) but in the Skaergaard it is clear that slowly cooled magnetite of the early stages is free from ex-solution ilmenite while later magnetite formed under similar conditions, although no doubt with a rather different composition, shows ex-solution of ilmenite.

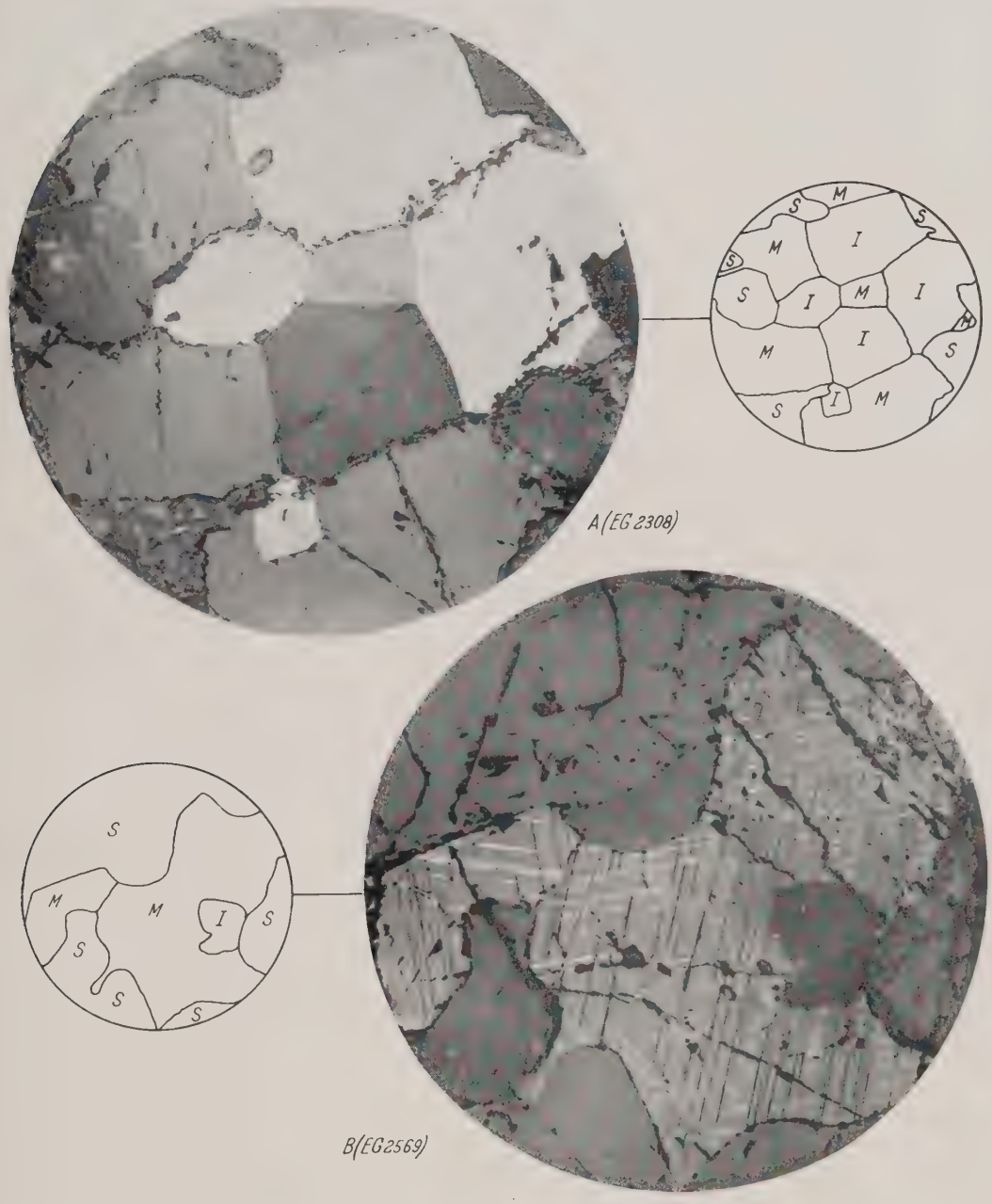


Fig. 4—Photomicrographs of polished specimens showing iron ores.

A. hypersthene livine gabbro, 2308, rich in iron ores; *M:* magnetite; *I:* ilmenite; *S:* silicates (see key).

B: hortonolite ferrogabbro, 2569, richer in iron ore but at about the horizon of the analysed rock 1907; *M:* magnetite with ex-solution lamellae; *I:* ilmenite; *S:* silicates (see key).

Polars 5° off cross position. Magnifications *A*: × 45. *B*: × 60. Photos by R. Phillips.

ilmenite structure should also be limited, whereas chromium is a major constituent of a considerable range of the spinels. In the second magnetite, separated from the Middle Gabbros (C stage) the amount of chromium is less than the sensitivity, *i.e.* < 1 ppm. Lundegardh (1945, 1946) has also encountered chromium-free magnetites in late differentiates. The iron of the Skaergaard intrusion is more reduced than is usual in basic magma, and this is probably the reason for the relatively late appearance of magnetites as primary precipitates. As has been suggested previously (1939, pp. 310–311) it is likely that magnetite would have separated at an earlier stage of the crystal fractionation if the magma had been more oxidized, and had this happened the early magnetites would have been richer in Cr.

Vanadium is present in both ilmenite and magnetite but the amount in magnetite is consistently two or three times the amount in ilmenite. Vanadium magnetites have been described from India (Dunn and Dey, 1937) and artificial spinels with V^{+++} as a major constituent have also been produced. It cannot be doubted that V is also present in the ilmenite in considerable amounts, as contamination cannot account for the values obtained. Thus contamination of the ilmenite from 4077 (column B) by sufficient magnetite to give 600 ppm of V is excluded by the low figure for Cr in the ilmenite. The role of V in ilmenite is not clear but it is presumably replacing Fe^{+++}, which can enter ilmenite to some extent.

Nickel and *Cobalt* enter both ilmenite and magnetite replacing Fe^{++}, and the amounts are roughly similar to the amounts found in corresponding pyroxenes and olivines. Thus Ni is abundant in early iron ores and is absent in the later ones, while Co persists throughout the series. As Lundegardh (1946) has suggested, the ratio Ni/Co may be used to indicate the stage of fractionation reached.

Copper is present in the magnetites and ilmenites, no doubt replacing ferrous iron. The amount increases with fractionation up to the ferrohortonolite ferrogabbro stage after which we have only a determination of copper in magnetite from the fayalite ferrogabbro and it is possible our sample is contaminated. In this rock pyrrhotite is present associated with chalcopyrite, and it may be that our method of magnetic separation has resulted in some pyrrhotite and chalcopyrite being included in the sample of magnetite analysed.

Zinc is present in the magnetites here analysed in amounts lying between 300 and 3000 ppm, whereas it is below the sensitivity, about 250 ppm, in the ilmenites. These figures are not given in the tables as their accuracy is not of the same order as for the others. Since Zn spinels are known, the presence of Zn in the magnetite is not unexpected but the amount is of interest. Neumann (1949, p. 80) has discussed the geochemistry of Zn and shows that it tends to replace Fe when in fourfold coordination. The presence of Zn in magnetite in which the Fe has fourfold coordination and its virtual absence in ilmenite in which the Fe has sixfold coordination is in harmony with Neumann's conclusions.

Molybdenum is present in both magnetite and ilmenite in small, and rather irregular amounts. An unusual magnetite having 400 ppm Mo has been described by Zeis (1938).

The trace element content of the iron ores is of considerable interest in relation to the fractionation process and it will become more so when further work allows comparisons with the iron ores of the Bushveld and other layered intrusions. Some comparisons have already been made (Wager and Mitchell, 1950, pp. 147–148). It has been pointed out that the Bushveld iron ores and titaniferous iron ores analysed by Landergren (1948) are richer in chromium and nickel than the earliest Skaergaard magnetite. Had magnetite separated earlier from the Skaergaard magma, when there was more chromium and nickel present in the magma, then it is to be expected that this early magnetite would have resembled in a general way the titaniferous Bushveld magnetites (Hall, 1932, Table XLIII, p. 346) and also the average titaniferous iron ores of north Sweden (Landergren, 1948, p. 108). As suggested above, precipitation of magnetite would no doubt have occurred earlier in the Skaergaard complex if the magma had been in a less reduced condition. On the other hand, the apatite iron ores of Kiruna are poor in chromium and are comparable with iron ores of the Skaergaard intrusion from the Middle Gabbros (column D). Shortly after the Middle Gabbro stage in the Skaergaard differentiation

apatite became a primary precipitate. The Skaergaard data suggest that apatite iron ores belong to a later stage of fractional crystallization differentiation than the typical titaniferous iron ores.

5. Apatite

Apatite is present only in minute amounts in the early rocks of the layered series, and there it has crystallized from the inter-precipitate magma. It becomes a primary precipitate mineral, however, 100 m above the analysed hortonolite ferrogabbro. 1907, and the rocks then contain 2 to 5% of apatite. Apatite was separated from one of the earlier rocks (3649) which contains it abundantly, and a small amount was also separated from the late fayalite ferrogabbro 4142 where the crystals were small and separation laborious. The results of the analyses are given in the last columns of Table B; the discussion is brief as the results of only two analyses are available.

Vanadium, *Molybdenum*, *Lithium*, *Nickel* and *Cobalt* are below the sensitivity in these two apatites and the amounts of *Gallium* and *Chromium* recorded are probably not significant. There is apparently no appreciable tendency for V to replace phosphorus in these apatites. This may be a somewhat unusual feature as the apatites became a primary precipitate late in the fractionation when the amount of V in the magma was already much reduced.

Copper is present in the apatites in amounts similar to that in the contemporaneous plagioclases.

Scandium and *Zirconium* are present in small amounts in both the apatites, presumably replacing Ca, as in the pyroxenes.

Manganese is present in both the apatites. The amount is very much less than that in the pyroxenes and olivines, but is greater than the amount in the contemporaneous plagioclases.

Yttrium and *Lanthanum* are abundant in both the analysed apatites. Rare earths replacing calcium are known to be often important constituents of apatites; for instance Kind (1938, p. 63) has shown that there is 1·49% of the rare earths in an apatite from an essexite and still more in apatites from syenites and nepheline syenites; rare earths are separated from the apatite associated with the nepheline syenites of Kola. In a series of apatites separated from the Caledonian igneous rocks of Scotland, Nockolds and Mitchell (1948, p. 568) show the presence of Y and La in somewhat similar amounts to the Skaergaard apatites. Y is about three times as abundant as La in the Skaergaard apatites and both elements show a significant increase during fractionation.

Strontium is abundant in both the apatites, presumably replacing Ca, while the amount of *Barium* is small. The evidence suggests that there is a fall in the Sr content with fractionation, like that which takes place in the contemporaneous plagioclases. The barium content is increased with fractionation as also occurs in the plagioclases.

Lead is at about the limit of sensitivity in the later apatite and below this in the earlier.

6. Trace Elements in Certain Minerals associated with Hybrid Rocks

In the border group and in the layered series, except for the upper part, granophyre patches occur with a coarse hybrid zone between the granophyre and surrounding gabbro (1939, pp. 185–188). Evidence has been given that these patches of granophyre are remelted blocks of acid gneiss which became incorporated in the original Skaergaard magma, presumably at the time of the initial intrusion. The hybrid material is coarse, pegmatite-like gabbro formed by some degree of mixing between the acid gneiss and the enclosing olivine gabbro. From the coarse hybrid zone large crystals of plagioclase and pyroxene grew into the granophyre mass (1939, Figures 37 and 38, and Plate 14, Figure 1). Certain minerals from these complicated hybrid

rocks have been examined spectrographically. As they are distinct in paragenesis from the primary precipitate minerals so far described, the data are presented separately in Table 7.

Table 7—Trace elements in minerals from the coarse hybrid zones surrounding granophyre inclusions, and in acid plagioclase and quartz from a miarolitic cavity in a granophyre inclusion

	r	s	Z	I Plagioclase 317	II Pyroxene 317	III Apatite 317	P	IV Plagioclase 1835	V Pyroxene 1835	VI Magnetite 1835	Q	VII Acid Plagioclase 4074	VIII Quartz 4073	R
				Trace constituents as parts per million of the element										
Ga^{+3}	0·62	1	25	40	5	*	20	50	8	30	50	20	*	40
Cr^{+3}	0·64	1	33	*	3	*	*	*	3	*	*		3	30
V^{+3}	0·65	5	28	*	150	*	20	*	80	1000	*	*	*	15
Mo^{+4}	0·68	1	*	*	*	*	*	*	*	10	*	*	*	*
Li^{+1}	0·78	1	17	*	8	*	2	*	5	2	3	*	2	1
Ni^{+2}	0·78	2	6	*	100	*	7	*	80	800	20	*	*	15
Co^{+2}	0·82	2	5	*	60	*	5	*	40	20	3	*	*	3
Cu^{+2}	C.0·8	10	C.15	20	300	*	20	*	*	*	20	*	*	60
Sc^{+3}	0·83	10	*	*	100	*	*	*	100	*	*	*	*	*
Zr^{+4}	0·89	10	160	*	50	10	500	*	80	*	+10,000	*	*	2000
Mn^{+2}	0·91	10	230	20	3000	200	—	20	3000	5000	—	20	*	300
Y^{+3}	1·06	30	C.30	*	200	C.10,000	200	*	500	*	200	*	*	800
La^{+3}	1·22	30	C.30	*	200	2000	300	*	300	*	200	*	*	500
Sr^{+2}	1·27	10	900	1000	10	800	100	1000	10	10	70	100	*	300
Pb^{+2}	1·32	10	*	*	25	*	*	*	*	*	*	*	*	*
Ba^{+2}	1·43	5	1800	200	*	*	500	600	*	10	10	30	*	1200
Rb^{+1}	1·49	20	160	*	*	*	30	*	*	*	*	*	*	80

Data determined spectrographically.
An asterisk indicates that the element is present in amounts less than the sensitivity.
A dash indicates absence of data.
The following elements were also sought but are not present in amounts above the sensitivity which is indicated in brackets : B(1), Be(10), Ge(10), Sn(5), Zn(300), Sb(300), Ce(500), Cd(300) Ag(1), Tl(50).
r Ionic radius in Angstrom units at valency shown.
s Sensitivity by the spectrographic method employed.
Z Average " Grey gneiss " (Table E, column B).
I Plagioclase C.An45
II Pyroxene
III Apatite
} from coarse hybrid material, 317, between granophyre, 318 (see B below), and olivine gabbro of western border group (1939, p. 191).
P Granophyre 318 within hybrid zone, 317 (see Table E, column I).
IV Plagioclase C.An45
V Pyroxene (γ'1·718)
VI Magnetite
} from coarse hybrid material, 1835, between granophyre 1836 (see Q below) and olivine gabbro, 1837 (see Table B, column V) of western border group (1939, p. 191).
Q Granophyre 1836 within hybrid zone 1835 [1939 Pl 14 Fig. 1]. (See Table E, column II).
VII Acid Plagioclase near Albite
VIII Quartz
} from miarolitic cavity of " granophyre " in the hypersthene-olivine gabbro (1939, pp. 185-186).
R " Granophyre " 4074 (1939, pp. 185).

Analyses I to VI (Table 7) are of minerals from the inner part of the hybrid zone where the crystals project into the central granophyre. Of these analyses I to III are from rock 317 which is shown in the lower part of Figure 37 in the 1939 paper and analyses IV to VI are from large crystals projecting into the central granophyre which is shown in Plate 14, Figure 1 in the 1939 paper. These crystals must have formed earlier than the time of solidification of the granophyre, and diffusion has probably brought material for their growth from some distance within the granophyre. The acid plagioclase and quartz analyses in columns VII and VIII are from single crystals projecting into a miarolitic cavity in the central part of a granophyre mass in the lower layered series (1939, pp. 185–187).

Plagioclase analyses I and IV have a trace element composition within the range already given for the layered series. The acid plagioclase VII is markedly different. This plagioclase is an albite formed at a low temperature, simultaneously with well shaped β-quartz, and it is noteworthy that the amount of strontium and barium is roughly a tenth of the amount found in the earlier plagioclases I and IV. Indication of a fall in the amount of strontium and barium in late felspars is given by Noll, v. Engelhardt, and Nockolds and Mitchell (1948, pp. 569-570). With regard to strontium, Noll (1934) finds that the strontium oxide content of adularia from St. Gotthard is 0·04%, while for plagioclase of a wide range of compositions he gives values from 0·05 to 0·1%. When considering barium, v. Engelhardt (1936, p. 216) found that potash felspar or perthite of late formation, has a very low content as, for example, the microcline from the Kragero pegmatites with 0·001–0·003% barium oxide, while early orthoclase and sanidine may contain up to 0·2% barium oxide.

The analysis of the quartz associated with the acid plagioclase in the drusy cavity, column VIII, was done as a blank test and it shows that most trace elements are below their sensitivity. The traces of chromium and lithium shown in this analysis are probably due to flakes of a brown micaceous mineral which was embedded in minute quantities in the outer surface of the quartz crystal.

In trace constituent composition, the two pyroxenes from the coarse hybrid zone (analyses II and V) are not unlike the pyroxenes of the middle gabbro, 3662 (Table B) except that both contain yttrium and lanthanum, a feature only shown by the latest analysed pyroxene in the layered series. It should be noted that the re-fused gneiss which formed a sort of flux for the formation of these pyroxenes, contains much more yttrium and lanthanum than the original Skaergaard magma (Table 7, columns B and C).

Apatite in the coarse hybrid rock occurs in crystals, 1 mm across and 2 cm long, which are grey with minute inclusions. This apatite is free from the trace elements for which we have analysed except for those of about the ionic size of calcium, namely Zr, Mn, Y, La and Sr. As in the apatites from the layered series yttrium is much more abundant than lanthanum and both are more abundant than strontium.

A magnetite from the hybrid rocks is not unlike in composition a magnetite from the middle gabbros, minor differences being that nickel is high relative to cobalt, and copper is relatively low.

The minerals considered in this section are formed partly from the gneiss inclusions, and the trace element composition of average Grey Gneiss is given in the table for comparison (column Z). This shows small amounts of most of the trace constituents with which we are here concerned. The trace element composition of the central mass of granophyre with which the minerals are associated is also given (columns P, Q and R). These rock compositions are commented upon later (pp. 169–173).

7. Summary and Discussion of the Results of the Mineral Analyses

The crystal chemical factors controlling the entry of trace elements into crystals have been finely elucidated by the work of Goldschmidt who gave various summaries of his views (1934, 1937, 1945); and it will be clear that the results given here provide further substantiation of his generalizations. Having noted that the trace elements present in minerals are, for the most part, not accidently incorporated, but that certain trace elements are associated with certain mineral species, he showed that the trace elements replacing a particular major element are not those related to it by the chemical relationships such as are expressed by the periodic classification but are those which have approximately similar ionic size, this being the chief factor governing the entry of a trace element into the crystal structure; he also showed that, if the ionic size is approximately right, a difference in ionic charge from the major element being replaced does not preclude the entry of a trace element, providing that other replacements allow the total charge on the crystal structure to be balanced. The tolerance in relation to size exhibited by ionic crystals clearly differs somewhat with the structure, but on the average, atoms within 10–15% of the ideal size may be found in the lattice. At higher temperatures atoms within 20–30% of

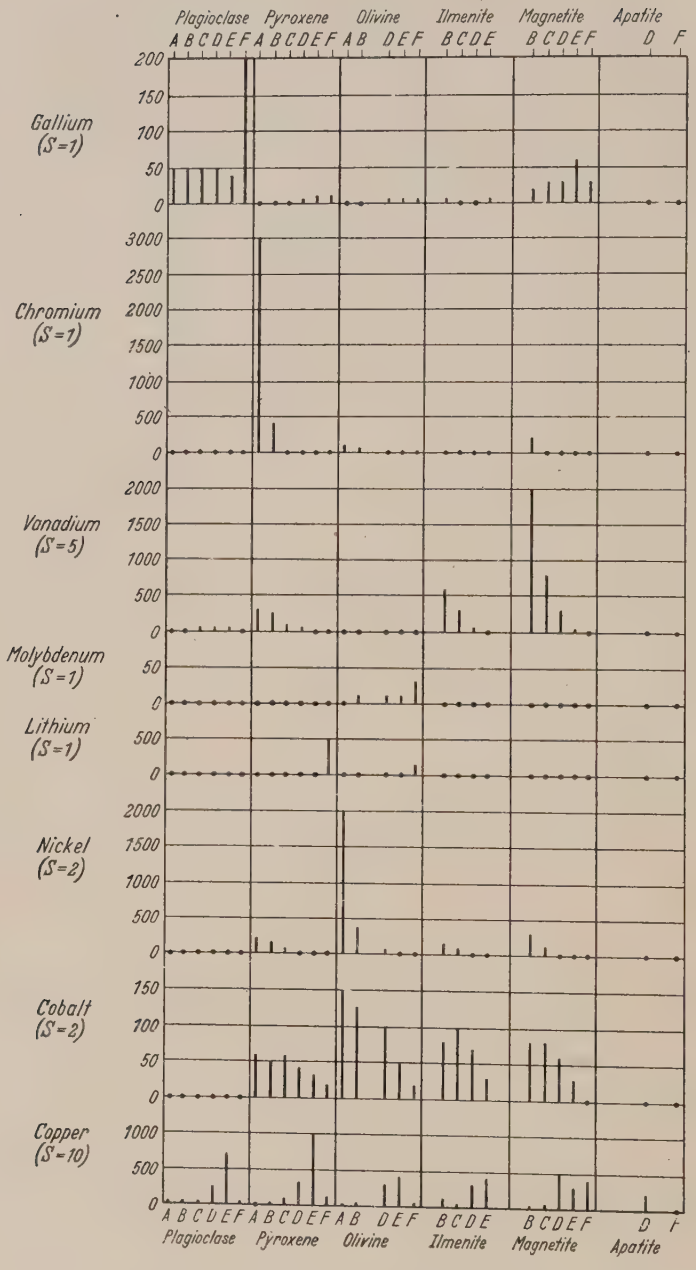
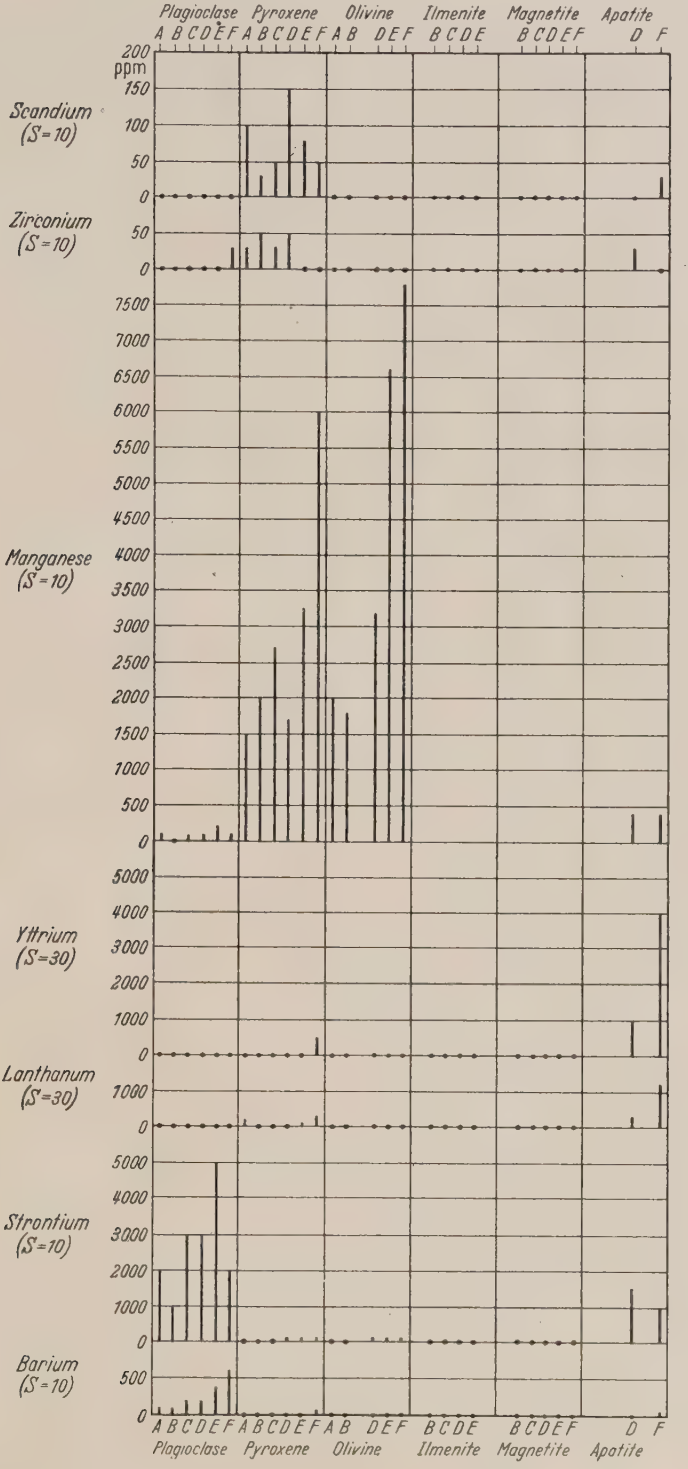


Fig. 5—Trace elements in the minerals of the layered series. Amounts are given as parts per million. A dot in any column indicates an amount below the sensitivity. Column *A:* minerals from early border rocks; *B:* minerals from hypersthene olivine gabbros; *C:* minerals from Middle gabbros; *D:* minerals from hortonolite ferrogabbros; *E:* minerals from ferro-hortonolite ferrogabbros; *F:* minerals from fayalite ferrogabbros.


Plagioclase
Pyroxene
Olivine
Ilmenite
Magnetite
Apatite
A B C D E F A B C D E F A B D E F B C D E B C D E F D F
ppm
Scandium (S=10)
Zirconium (S=10)
Manganese (S=10)
Yttrium (S=30)
Lanthanum (S=30)
Strontium (S=10)
Barium (S=10)
200
150
100
50
0
50
0
7500
7000
6500
6000
5500
5000
4500
4000
3500
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A B C D E F A B C D E F A B D E F B C D E B C D E F D F
Plagioclase
Pyroxene
Olivine
Ilmenite
Magnetite
Apatite

the ideal size may enter, but such crystals are liable to undergo ex-solution. Besides size and charge, another factor visualized as polarization of the ions, or a partial approach to homopolar bonding, is also involved. A full evaluation of the tendency of a trace element to enter any particular crystal structure would necessitate a quantitative treatment of these and other concepts and it will be long before the observational approach is superseded.

While Goldschmidt verified these generalizations on the entry of trace elements into crystals by a study of the trace element composition of minerals from many different localities, the data here presented are based on the study of series of genetically related minerals from one intrusion. These data are summarized in the successive diagrams of Figure 5, which show, in the first place, the minerals in which a particular trace element is present, and, in the second place, the variation in the amount of the trace elements in the various minerals of successive differentiates. The diagram may be compared with a similar one for minerals from Caledonian plutonic rocks (Nockolds and Mitchell, 1948, Figure 9) and, when comparison is possible, the same general results are apparent. An important point about the Skaergaard data is that the figures in corresponding columns are for minerals which were formed simultaneously, as relatively unzoned crystals, from a well stirred liquid. This allows conclusions to be drawn without the extensive averaging of data which is necessary when comparisons are made between minerals from various rock bodies.

The general distribution of the various trace elements in the several minerals is sufficiently shown in the diagram, but it is desirable to accompany this by an indication of the probable situation of the trace elements in the crystal structures. The following are the probable replacements:

Ga for Al in plagioclase, and for Fe^{+++} in magnetite.

Cr for Fe^{+++} in pyroxene and magnetite.

V for Fe^{+++} in pyroxene, magnetite and ilmenite, to some extent, and probably for Al^{+++} in plagioclase, to a small extent.

Fe^{+++} for Al in the silicon-oxygen tetrahedra of plagioclase and perhaps of other minerals.

Mo apparently for Si in olivine.

Li for Mg and Fe^{++} in late pyroxene and olivine.

Ni, Co and Mn for Mg and Fe^{++} in pyroxene, olivine, ilmenite and magnetite.

Sc, Zr and probably Mn for Ca in pyroxene and apatite.

Y and La for Ca in apatite and in late pyroxene but not appreciably in plagioclase.

Sr for Ca in plagioclase and apatite but scarcely at all for Ca in pyroxene.

Ba for Ca in plagioclase but scarcely at all in apatite.

The elements replacing Al or Si within the oxygen tetrahedra seem to be Ga, Fe^{+++} and perhaps Mo, Cr and V. The replacement of Si within the oxygen tetrahedra by Ti has been shown for garnets and probably for pyroxene (*cf* Barth, 1931, and Wager and Deer, 1938) but the extent of the other replacements suggested here deserves further investigation.

With regard to the replacements of the cations outside the silicon oxygen tetrahedra, there is a general point which should be emphasized by means of the present data, although it has been made before, for instance by Wickman (1943, p. 383). An element may replace another readily in one mineral, but not this same element in another mineral forming simultaneously. Strontium for example, replaces calcium in plagioclase, but it replaces to only a slight extent the calcium of diopsidic pyroxene. It is clearly better in this case not to speak of the replacement

of one particular element by another but rather to say that strontium can occupy the calcium positions in plagioclases, the structure adjusting to take it, while it cannot occupy the calcium positions in pyroxenes to an appreciable extent, the structure, apparently, not being able to make the necessary adjustments. Another example from our data is the replacement of Ba for Ca in the plagioclase, but scarcely at all for Ca in apatite separating simultaneously. A similar case, but involving valency adjustments is that of Y and La which enter apatite and late pyroxene, presumably replacing Ca, while they are not able to replace appreciably the Ca in the plagioclase structure. Still other examples are given in Nockolds and Mitchell (1948, pp. 552 and 554).

When replacements of elements of different valencies take place adjustments have to be made by the complementary introduction elsewhere in the lattice of an element with a different valency to produce general neutrality over the lattice. Here again the flexibility of the whole structure is the important factor and it is often undesirable to think of the replacement of certain specific elements by others. The series of diopside-hedenbergite pyroxenes here analysed can accommodate elements with radii ranging from Al 0·57Å to K 1·33A and with various valencies, for example, Li, Na, K of valency 1; Ni, Mg, Co, Cu, Fe, Mn of valency 2; Cr, V, Fe, Sc, Y, La of valency 3; and Zr, Ti of valency 4; and these elements are accommodated in the two situations normally occupied by Ca and Fe/Mg. It is obvious that, in attempts to predict the relative amounts of trace elements entering a complex crystal like pyroxene, it will be necessary to go far beyond the simple concept of replacements in terms of size and charge, but discussion of these matters is beyond our scope.

Considerations of the size, charge, *etc* of ions, allows predictions to be made of the elements which may enter a particular structure and which may not. Without petrogenetic knowledge such crystal-chemical concepts do not, however, indicate which of the possible elements will actually be present in any particular mineral species as developed in the earth. Goldschmidt proceeded to consider this latter, more purely geological, problem. Believing that fractional crystallization of magma is one of the chief factors in producing the varied igneous rocks of the earth's crust, he considered how trace elements would enter different crystal fractions. He showed that those entering abundantly the early crystal fractions should be elements of smaller ionic size, or higher ionic charge, than the replaced ions, or both. These elements he described as *captured* by the host crystal, by virtue of size or charge. From among the trace elements capable of entering a crystal structure, those which enter the later crystal fractions more abundantly are those of large ionic size or of lower ionic charge than the element replaced; these are *admitted*, to use Goldschmidt's term, only when the supply of more appropriately sized elements is reduced. The processes involved in trace element capture by a host crystal, on the one hand, and in late acceptance, on the other hand, are essentially the same as those controlling the major element variation in solid-solution series during fractionation. Goldschmidt illustrated this briefly in relation to size and charge by reference to the forsterite-fayalite series and the anorthite-albite series (1937, p. 660). These effects can be seen very clearly at work on both the major and trace constituents of the various mineral series developed by fractionation in the Skaergaard intrusion.

The distribution factor between crystal and liquid is of some value in estimating the tendency of a particular element to enter into a crystal. It is a device whereby

the effect of the concentration of the element in the liquid is eliminated and in using it we are assuming that an increase or decrease in the concentration of the trace element in the liquid will produce a proportionate increase or decrease in the amount in the crystal phase. For a major constituent, variations in the composition of the magma, even when not involving the element in question, may have a big effect on the distribution factor for the element; and indeed, the distribution factors of major constituents are only of interest in relation to the values for related trace elements. On the other hand, for the trace elements, variation in the composition of the liquid, other than in the amount of the trace element in question, probably has little effect on the distribution factor. Consideration of distribution factors emphasizes the importance of the concentration of the trace element in the liquid in controlling the amount of the trace element in the crystal phases separating.

In attempting to make use of distribution factors in this investigation we unfortunately find considerable irregularities due to the inaccuracy of some of the data (Table 6). Thus, nickel enters with about equal ease into the olivines of the A, B and D stages, judging by the distribution factors of about 10, and it is likely that nickel also enters with about equal ease into the A, B, C and D stages of the pyroxenes, although the distribution factor shows considerable variation. The low distribution factor of 1·2 for the first pyroxene (column A) is probably due to the underestimation of the nickel in the pyroxene crystals, and the decline in the distribution factors for later pyroxenes and olivines is probably due to the poor estimates of the nickel content of the later liquids when the amount falls to low values of about 1 ppm or less. Since the values for the amounts of trace elements in the crystal are subject to considerable errors, and the estimation of the composition of the liquid is a further possible source of error, the estimated distribution factors given in Table 6 are, in many cases, unreliable. Distribution factors believed to be misleading have been put in brackets.

In a magma which crystallizes without fractionation (except that due to zoning of the crystals), a low amount of an element in a crystal is directly the result of difficulty of entry into the crystal structure, and a high amount is directly the result of ease of entry. The trace elements from a small portion of the magma are all sorted into the crystal phases formed from this small unit of magma and the effects resulting from fractionation are absent. With fractionation such as gave the minerals of the Skaergaard layered series, the factors are more complex. A low amount in a crystal may be due to: either, difficulty of entry, *e.g.* Fe^{+++}, Cr^{+++}, Fe^{++}, Ni^{++} in plagioclase, Sr in pyroxene, *etc*; in these cases the distribution factors are all low, or to ease of entry into early crystals which are precipitated in sufficient abundance to remove so much of the element from the magma that the later crystals, precipitating from the impoverished magma contain necessarily little or none of the element, *e.g.* Cr, V and Ni in late pyroxenes and iron ores.

Similarly, a high amount in a crystal may be due to: either ease of entry, *e.g.* Cr in early pyroxene, Ni in early pyroxene, olivine, *etc*; in these cases the distribution factors are all high; or to difficulty of entry into all phases in the early stages of fractionation, so that the amount in the liquid increases and that, in the crystals increases because of the increase in the amount in the liquid, *e.g.* Ba in plagioclase; Y and La in pyroxene; Li in pyroxene and olivine.

In some cases there may be difficulty of entry in the early stages, leading to an increase in the amount in the liquid, then, in the later stages, increased ease of entry, resulting in a greatly increased amount in the crystal phases at a middle stage. Thus in the case of copper, the distribution factors for the early plagioclase, pyroxene

and olivine are only about 0·1, and this leads to a steady increase in the copper content of the liquid over the first half of the differentiation process. However, the distribution factors for these three minerals, and also for ilmenite and magnetite, when they become primary precipitates, increase until they reach 1 or 2. This results in heavy precipitation of copper in the various silicate minerals, a process apparently brought to an end by the formation of an immiscible liquid sulphide phase.

The relative changes in amount of the various trace elements in a series of minerals resulting from fractionation are the result of an interplay between the crystal chemical factors such as size, charge and polarization of the trace element, on the one hand, and the sequence and amount of the various crystal phases produced during the crystal fractionation on the other hand. The sequence and amount of the crystal phases are the result of complicated thermal equilibria which are controlled by the major components of the system, and these equilibria are, of course, only imperfectly understood. Nevertheless an attempt will be made here to consider summarily the way in which the general principles given in the preceding paragraphs can be used to explain the changes in the relative amounts of trace elements in the mineral series of the Skaergaard intrusion. The pyroxenes, olivines and iron ores will be considered first and then the plagioclases.

For the pyroxenes, olivines and iron ores the decrease with fractionation in the amounts of chromium, vanadium and nickel is due to markedly easy entry into these crystals, as shown by their high distribution factors, and to abundant precipitation of the crystal phases containing them, so that the amount in the liquid rapidly decreases. (It is not believed that there is a lowering of the otherwise high values of the distribution factors in the later stages of the fractionation, but that the low values for the distribution factors given in brackets in Table 6, are unsatisfactory because of the difficulties in estimating the low amounts in both the crystals and the liquid.) For the same group of minerals, the increase in the amounts of lithium, copper, manganese, with fractionation, is due, firstly, to a low ease of entry into the various phases separating, resulting in accumulation of these elements in the magma, and secondly, to an increase in the ease of entry in the later stages, as indicated by the distribution factors (see Table 6). Again for these same minerals the moderate amount of cobalt in the early stages and the small decrease with fractionation is due to moderate entry into the early phases so that the cobalt content of the magma only falls slowly. There is an indication from the distribution factors that the ease of entry of cobalt may be slightly greater in the middle and later stages, and this tends to maintain the cobalt figures over the middle range and leads to a slight accentuation of the fall in the later stages. The case of scandium is somewhat similar; it enters readily into the pyroxene but not into the other minerals; the amount of pyroxene precipitated is such that the scandium content of the liquid falls only slowly and at the same time the ease of entry into the pyroxene as judged by distribution factors, increases, with the net result that there is at first a slow rise in scandium content and then a fall. The results for zirconium are more unexpected. Among early minerals, zirconium only enters pyroxene and the distribution factor is about 0·5; thus zirconium accumulates in the magma but for some reason the distribution factor for the zirconium in pyroxene falls rapidly so that the actual amount in the pyroxene also falls, despite the increasing zirconium content of the liquid. The zirconium in the magma continues to rise until eventually a point is reached when zircon is precipitated as a separate crystal phase.

For the plagioclases, the very slight increase in the amount of potassium, despite a fairly rapid increase in the amount in the liquid, is apparently due to reduction in the ease of entry of the K ion. The distribution factors confirm this as they change from 1·1 to 0·6 over the short range from stages B to E. Barium on the other hand, shows a twentyfold increase in amount during fractionation, although the ease of entry, as judged by the distribution factor, remains essentially constant. Barium is extracted in such low amounts by the minerals separating in the early stages that it accumulates rapidly in the magma and the high amounts in the later plagioclase are proportional to the increased amount in the magma. Strontium, on the other hand, enters fairly abundantly into the early plagioclase and scarcely at all into the pyroxene or other minerals. Conditions happen to be such that the amount in the magma varies little. The distribution factor seems to rise until near the end of the series and then falls; the amount of strontium in the crystals shows the same behaviour because of the relatively constant content of strontium in the magma.

In attempting to understand the amounts of trace elements in a series of minerals produced by fractionation, it is clear that petrological factors, such as the sequence and quantity of the crystal phases precipitating, must be considered as well as the relevant crystal-chemical factors, a point which we have already briefly discussed (Wager and Mitchell, 1950, p. 148). In particular the varying amounts of the trace elements in the successive residual liquids must be taken into consideration when attempting to understand the trend in the amounts of the trace constituents in mineral series produced by fractionation, and this is a complex matter dependent on the summation of the effects due to the sequence and amounts of the crystal phases successively separating. Since the trace element composition of a mineral depends not only on certain crystal-chemical factors, but also on what may be called petrogenetic factors, the amounts of trace elements in minerals will no doubt become a valuable clue in the study of petrogenesis, but before this is possible there will have to be many studies of the trace element composition of minerals from well defined rock series whose genesis is reasonably well understood.

VII. The Trace Element Composition of the Rocks of the Intrusion

The rhythmic banding of the layered series produces strong variations in the proportions of the minerals and this naturally produces strong variation in the amount of the major and trace constituents. We have determined the trace constituents in two strongly contrasted bands at the hortonolite ferrogabbro horizon, a felspathic band 2568 and a melanocratic band 2569 consisting largely of olivine, pyroxene and iron ore. The average rock at this horizon is the ferrogabbro 1907 already investigated. The results, Table 8, show strong differences in the trace elements which are as would be anticipated from the mineral proportions, and the known amount of trace constituents in the minerals. In discussing the fundamental trend of differentiation of the Skaergaard magma, and of many other magmas also, it is necessary to concentrate attention on the composition either of the individual minerals or of the average rocks.

Trace element data for the average layered rocks, whose modal composition is shown in Figure 3, are presented in full in Table C, columns VII–XIX. The order is that of increasing height in the intrusion. The rocks of columns VIIa to XVI have been shown by the earlier work to consist of successive crystal fractions, separated

out, with great regularity, at the bottom of a pool of convecting magma. The unlaminated layered series, columns XVII to XIX, marks a break in this regular development, the rocks being interpreted as the result of solidification of a layer of residual magma undisturbed by convection; the upper part is basic granophyre in composition (columns XVIII and XIX) while the low layer (column XVII) is essentially the same material with the addition of large fayalite crystals which have

Table 8—Trace elements in contrasted bands at the hortonolite ferrogabbro horizon

	s	*Leucocratic band* 2568 SG 2·75	*Average hortonolite ferrogabbro* 1907 SG 3·18	*Melanocratic band* 2569 SG 3·74
P			350	
Ga	1	30	20	10
Cr	1	*	*	*
V	5	20	15	80
Mo	1	*	*	2
Li	1	3	3	*
Ni	2	*	*	3
Co	2	7	40	60
Cu	10	200	400	400
Sc	10	*	10	*
Zr	10	100	20	20
Mn	10	500	(1300)	3000
Y	30	*	*	*
La	30	*	*	*
Sr	10	1000	700	300
Ba	5	150	50	*
Rb	20	*	*	*

An asterisk indicates that the element was present in amounts below the sensitivity.

apparently sunk into it from above. So far as could be judged in the field these analyses represent average rock. Subsequent comparisons confirm this average character, except that 4076 (column VIIB) is rather melanocratic; 3662 (column VIII) is rather plagioclase-rich, and 1881 (column XVI) has suffered from late stage hydro-thermal effects more than others of that horizon and is also considered to be transitional to the basic hedenbergite granophyres of the unlaminated layered series. Despite some irregularities due possibly to the variations inherent in the spectrographic technique employed and to errors of sampling, the data are satisfactorily consistent and the general trend can be shown by averages for the standard stages of the differentiation series. Columns not separated by vertical lines in Table C, contain the analyses of rocks of essentially the same stage, and averages of these are presented in Table F.

At a late stage in the differentiation other granophyres more acid than those of the unlaminated layered series have been produced in small amounts. These rocks are more irregularly variable than the layered series and a number of analyses have had to be made in order to obtain a fair conception of their composition (Table D). On the basis of occurrence and petrology they were divided into two groups, the transgressive hedenbergite granophyres and the acid granophyres. This grouping also fits the spectrographic data satisfactorily and averages of these two groups are given in Table F (columns H and I).

Certain rocks of the marginal and upper border groups have also been analysed spectrographically. Rocks of the marginal border group, the analyses of which are given in the first part of Table C, represent early rocks formed prior to any of the layered rocks at present exposed. The first two analyses are of gabbropicrite which has resulted from an especial concentration of olivine and chromite. The chromite was not described in the original memoir; it is present only in small amounts as rare groups of minute, cuboid crystals showing a synneusis arrangement and it was overlooked until spectrographic analyses were made. The two rocks, 1851 and 1846, are outer border olivine gabbros or eucrites and the previous petrological investigations suggested that they are solidified early magma with the addition, in small amounts, of certain crystal fractions in existence at the time. Two other border rocks analysed (1837 and 4289) are from a position somewhat further within the intrusion

Table 9—Trace elements in rocks of the upper border group and comparisons

	r	s	*A*	*I* 3052	*II* 3050	*B*	*III* 4163	*Z*
				Trace constituents as parts per million of the element				
P^{+5}	0·34		1260	(4000)	(1700)	3500	(9500)	(1700)
Ga^{+3}	0·62	1	20	30	30	21	20	25
Cr^{+3}	0·64	1	75	30	20	0·7	*	33
V^{+3}	0·65	5	220	300	400	10	30	28
Mo^{+4}	0·68	1	*	*	*	0·1	*	*
Li^{+1}	0·78	1	4·5	15	10	5	10	17
Ni^{+2}	0·78	2	50	40	20	0·6	*	6
Co^{+2}	0·82	2	35	40	20	22	15	5
Cu^{+2}	C.0·8	10	220	300	300	360	*	C.15
Sc^{+3}	0·83	10	15	10	10	9	10	*
Zr^{+4}	0·89	10	80	100	100	90	100	160
Mn^{+2}	0·91	10	1220	(1900)	(1600)	1500	(900)	150
Y^{+3}	1·06	30	13	*	*	45	60	C.30
La^{+3}	1·22	30	7	*	*	23	*	C.30
Sr^{+2}	1·27	10	500	500	700	470	400	900
Ba^{+2}	1·43	5	90	70	200	170	600	1800
Rb^{+1}	1·49	20	3	*	*	10	20	160

Data for trace elements determined spectrographically except where figure is in brackets when the values given are from chemical analysis.

An asterisk indicates that the element was present in amounts less than the sensitivity.

A dash indicates absence of data.

The following elements were also sought but are not present in amounts above the sensitivity which is indicated in brackets: B(1), Be(10), Ge(10), Sn(5), Zn(300), Sb(300), Ce(500), Cd(300), Ag(1), Pb(10), Tl(50).

r Ionic radius in Angstrom units at valency shown.

s Sensitivity by the spectrographic method employed.

A Estimated composition of the second liquid (see Table 10).

I 3052 Quartz gabbro, summit of Brodretoppen (1939, pp. 170–172). Chemical analysis gave:—ZrO_2=0·01% *i.e.* Zr=75 ppm, SrO=0·12%, *i.e.* Sr=1000 ppm, BaO=0·04%, *i.e.* Ba=360 ppm, NiO=0·01, *i.e.* Ni=80 ppm).

II 3050 Quartz gabbro, 200 m below summit of Brodretoppen (1939, pp. 70–72).

B Estimated composition of the third liquid (see Table 10).

III 4163 " Acid quartz gabbro " 450 m below summit of Osttoppen on north face (1939, pp. 170–173).

Z Average " Grey gneiss " (Table E, column Z).

and are of rather later consolidation. They have abundant olivine and rather small amounts of pyroxene with a patchy distribution and were described as approaching troctolite in the early paper (1939, p. 153). The small amount of pyroxene, especially in 4289, is reflected in the low chromium and vanadium of the analysis. These two border rocks were also mainly formed by solidification of a residual magma, estimated to be contemporaneous with the formation of the lowest accessible layered rocks.

The marginal border rocks being essentially of solidified magma have a different chemical composition from the early layered rocks which are composed mainly of the crystal fractions. Averages for the first two stages of the marginal border rocks are given in Table F (columns A and A′).

Three quartz gabbros of the upper border group which have previously been described and analysed for major constituents (1939, pp. 169–177) have been

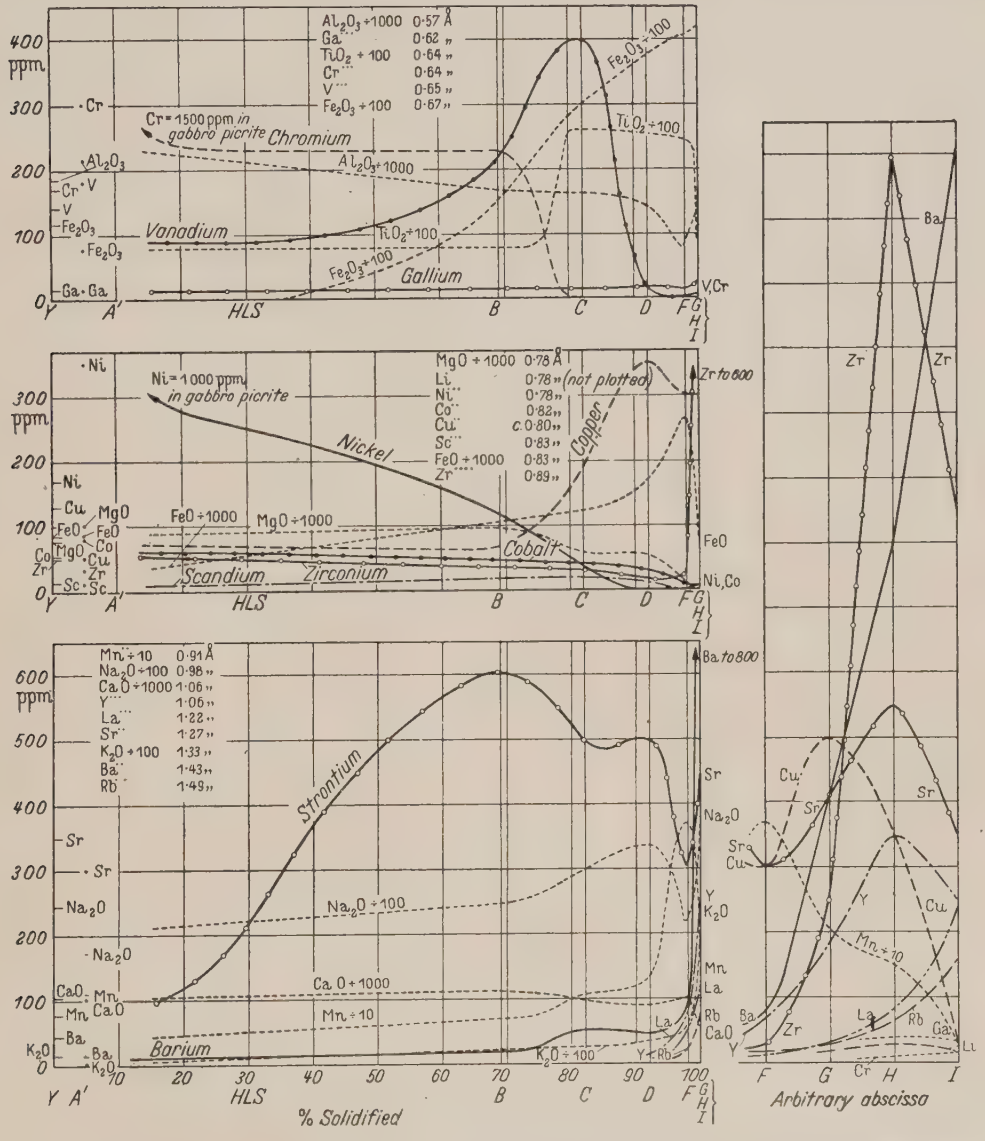

Fig. 6—Variation diagram showing the amounts of the elements in the rocks of the Skaergaard intrusion. *Y:* marginal olivine gabbro (the original magma); *A′:* early olivine gabbro of border group; *HLS:* estimated composition of hidden layered series; *B:* hypersthene olivine gabbro; *C:* Middle Gabbro (without olivine); *D:* hortonolite ferrogabbro; *F:* fayaliteferro gabbro; *G:* basic hedenbergite granophyre; *H:* hedenbergite granophyre; *I:* acid granophyre.

analysed spectrographically (Table 9). The upper border rocks are considered to be the result of solidification of successive residual magmas contaminated by the dominant " grey gneiss " of the metamorphic complex surrounding the intrusion. The two quartz gabbros 3052 and 3050 are considered to have formed roughly contemporaneously with the hypersthene-olivine gabbro 4077 (Figure 1). The composition of the liquid at this stage has been estimated (second liquid, Table 10) and the figures are quoted in Table 9 for comparison. The trace element composition of the somewhat later upper border quartz gabbro, 4143, from below 3052 and 3050, lies, on the whole, between compositions of the third and fourth liquids. The average composition of grey gneiss (column Z) differs from the Skaergaard liquids most markedly in the higher Ba and Rb content and rather higher Li content. In their high lithium and in some cases in their high barium content the upper border rocks perhaps show the effect of contamination with the grey gneiss.

The average figures for the amount of the trace elements in the rocks of the layered series and the later differentiates (Table F) are plotted against the percentage of the magma solidified at the time of formation of the rocks in Figure 6 (right hand part). Essentially similar curves would have been obtained if the values had been plotted against the composition of the plagioclase or against the iron-magnesium ratios. If plotted against silica percentages the curves would have been meaningless until the latest stages had been reached. The variation in major constituents are also given, the curves being taken from the previous paper (1939, Plate 27). A sympathetic relationship between trace and major elements is often to be seen. Thus the curves of gallium and aluminium show a sympathetic relation, so do the curves of nickel and cobalt on the one hand, and magnesium on the other; of vanadium, copper, manganese on the one hand and ferrous iron on the other ; of strontium and calcium; and of barium and potassium.

The curves showing the trace element composition of the successive rocks are a function of the proportion and trace element composition of the primary precipitate minerals, and of the composition of the interprecipitate magma which at one stage occupied the interstices. This will be discussed when the behaviour of each trace element is considered in detail later. Here the broad trends of related groups of curves may be briefly stated. Cr, Ni and Co all show falling curves although at different rates, except that there is a slight increase in chromium and nickel in the latest rocks. The curves for V, Cu, Sc and Sr reach various maxima and then fall. The curves for Ba, Y, La and Rb rise steadily to begin with and rapidly in the later stages. The curve for Zr descends slowly over most of the range, then rises very rapidly like the curve for Ba but finally falls unlike the Ba curve.

VIII. The Trace Element Composition of the Metamorphic Complex and of certain Granophyre Inclusions

The Skaergaard intrusion occupies a portion of the earth's crust formerly composed of the rocks of the metamorphic complex and overlying basalts. The previous work showed that some masses of gneiss became involved in the basic magma. Only those which were not completely incorporated can be identified, and they form granophyre inclusions and masses of hybrid material (p. 136 and p. 159). Some acid gneiss must have become completely incorporated, but during the previous investigations no satisfactory means was devised of estimating the amount; as a working hypothesis it was suggested that 1% by volume of the total intrusion was produced from complete digestion of acid gneiss (1939, pp. 292–293).

Several examples of the dominant grey gneiss, of quartz diorite and granodiorite composition, have been analysed, the specimens being selected from widely separated points in the Kangerdlugssuaq area (Table E, columns VII–XII). On the whole the range of variation is less than might have been anticipated for a banded metamorphic complex and an average can fairly be made which represents the composition of the major unit of the metamorphic complex, estimated as about four fifths of the whole (column Z). Rankama (1946) has collected together in tabular form the available figures for the trace constituents of granites and the acid parts of Pre-Cambrian complexes. There are also available figures for granitic rocks in the papers of Nockolds and Mitchell (1948) and Lundegårdh (1946). These data show so much variation that an attempt at this stage to make a detailed comparison between them and the results for the grey gneiss of the Kangerdlugssuaq area does seem worth while.

Irregular pink schlieren, richer in potash felspar than the normal grey gneiss, are not uncommon in the area and are considered to represent pegmatites now partly granulitized; one of these has been analysed (column XII). This single analysis is not sufficient to show significant differences in trace element composition from the bulk of the grey gneiss.

Two amphibolite bands of the metamorphic complex which have also been analysed show, as might be expected, an approach to the trace element composition of basic magma. One of these, the amphibolite 2013 (column XIV), contains less ferromagnesian minerals, and the plagioclase is more acid, than is the case with the second amphibolite, 341B (column XV) and its lower chromium and nickel and higher barium are in harmony with its more acid composition.

From a crude estimate of the proportions of grey gneiss and amphibolite given previously (Wager, 1947, pp. 9–10) and ignoring the small amounts of ultrabasic and sedimentary rocks incorporated in the complex, an estimate is given under V of the trace elements in the metamorphic complex as a whole. These figures should give some indication of the trace element composition of the upper crust in the area.

The granophyre inclusions in the Skaergaard intrusion which are regarded as having been formed from incorporated blocks of acid gneiss, have been described in some detail in the original paper (1939, pp. 185–199). Six examples have been analysed spectrographically, and there also exist data on the alkali content for four of these (1939, Table XXX). The analyses show considerable similarity among themselves and an average is given under U. In several respects the trace element composition of the granophyre inclusions differs significantly from that of the grey gneiss. Thus in the granophyre inclusions there is less Li, markedly less Sr, Ba and Rb, and markedly more Zr, Y and La. The number of rocks investigated is small, but these differences are accepted as significant although no convincing reason for the differences can be given. Two hypotheses may be advanced; first, that the material remaining as granophyre inclusions consists of grey gneiss of special composition which made it resistant to complete incorporation: second, that there were additions and subtractions of trace elements during the long period of cooling of the intrusion. The grey gneiss xenoliths must have remained, for a long period, as mainly fluid, viscous masses surrounded by a partly crystallized basic material, consisting of a crystal network with a residual liquid in the interstices, and they would be expected to undergo changes in composition, by material migrating into, and out of them. The factors involved in such migrations would no doubt be

complex; the few analyses of minerals from the granophyres and hybrids already presented are not sufficient to throw much light on the problem and we shall content ourselves, at present, with recording the trace element analyses without further comment.

IX. The estimated Trace Element Composition of the Successive Residual Magmas and of the Hidden Layered series

The mapping and previous petrological investigation of the Skaergaard intrusion allowed the original volume of the intrusion to be obtained and also the volume of the various component rocks. Using the known time sequence of the differentiates and assuming that the whole intrusion formed essentially a closed system, the composition of successive residual magmas may be calculated. This was done for the major constituents in the earlier paper (1939, pp. 217–224) and, using the same method, the composition of successive residual liquids in terms of trace elements is calculated here (Table 10). The acid granophyre I (Table F) is taken as the composition of the latest liquid, arbitrarily called the sixth, and assumed to be $\frac{1}{3}\%$ by volume of the intrusion. The fifth liquid, $1\frac{1}{2}\%$ of the intrusion, is estimated to consist of the sixth liquid together with $\frac{1}{3}\%$ of material of composition H, $\frac{1}{3}\%$ of composition G and $\frac{1}{2}\%$ of the composition of the fayalite ferrogabbro of the unlaminated layered series 4139 (Table C). From the figures in the second and third columns, which are proportional to the actual weights* of the elements in the residual magma, the amounts in the fifth liquid are obtained (fourth column), and finally the composition in parts per million of this liquid may be calculated (fifth column). The compositions of the residual magmas at various earlier stages are calculated in the same way. The so-called second liquid, amounting to 40% of the total intrusion, represents the magma at the time of formation of the earliest exposed layered rock—the hypersthene olivine gabbros. The composition of the original liquid (first liquid) is, as already explained, taken as the average composition of the chilled marginal gabbros (Section III). By subtracting the estimated amount of any trace element present in the exposed 40% of the intrusion, from the total quantity present in the original body of magma, the amount in the 60% of the hidden layered series (the part below the present level of erosion) is obtained and from this may be calculated the average trace element composition in parts per million of the hidden layered series (last column of Table 10).

After making these successive estimates, depending on the average compositions of a rhythmically layered intrusion, and on volume estimates based on ordinary field mapping, it is perhaps remarkable that the calculated average composition of the hidden layered series works out to be roughly what would be expected. The estimated composition of the average hidden layered series is fairly similar to the composition of early marginal border rocks (Table 11, columns II and III) and this is what is anticipated from the previous work. Also the figures for the average early marginal border group and for the hidden layered series differ from the composition of the earliest exposed layered rocks in the same direction, where the differences are considerable, except in the case of manganese (Table 11, columns II, III and IV). Comparing the composition of the original magma and the estimated composition of the hidden layered series (columns I and III of Table 11), it appears that the early layered series, amounting to about half the intrusion, had Cr and Ni contents about $1\frac{1}{2}$ to 2 times that of the original magma; Ga, Co, Sc and Zr contents

* If the figures given in the table, other than the percentages, are multiplied by 10,000 the amounts of the various elements will be obtained in metric tons.

Table 10—The estimated compositions of succession residual magmas and of the hidden layered series

(Figures for compositions are in ppm)

	Composition of fifth liquid I	1% of mean of grano-phyres G, H and I	½% fayalite ferrogabbro 4139	Amounts in 1½% fifth liquid	Composition of fifth liquid	1% fayalite ferrogabbro F	2½% ferro-hortonolite ferrogabbro 4145	Amounts in 5% fourth liquid	Composition fourth liquid	6% Horton-olite ferro-gabbro D	1% Upper border gabbro 4163	Amounts in 12% third liquid	Composition third liquid	9% Middle gabbro C	12% hypersthene olivine gabbro B	7% Upper border gabbros 3052–3050	Amounts in 40% second liquid	Composition second liquid	Composition first liquid (chilled marginal gabbro Y)	60% Hidden Layered Series	Composition Hidden Layered Series	
P	0	20	20	40	**2600**	70	200	310	**6200**	21	95	426	**3500**	26	31	20	503	**1260**	**280**	—	—	P
Ga	30	0·3	0·15	0·45	**30**	0·17	0·5	1·12	**23**	1·2	0·2	2·52	**21**	1·4	2·3	2·1	8·3	**20**	**17**	8·7	**14**	Ga
Cr	44	0·08	—	0·08	**5**	0·01	—	0·09	**1·8**	—	–	0·09	**0·7**	—	28	1·7	30	**75**	**170**	140	**230**	Cr
V	18	0·06	—	0·06	**4**	—	—	0·06	**1·2**	0·8	0·3	1·16	**10**	36	26	25	88	**220**	**140**	52	**90**	V
Mo	1	0·01	—	0·01	**1**	—	—	0·01	**0·2**	—	—	0·01	**0·1**	—	—	—	0·01	**0·02**	*	—	—	Mo
Li	13	0·2	0·05	0·25	**17**	0·06	0·05	0·36	**7·2**	0·12	0·1	0·58	**5**	0·27	0·2	0·8	1·8	**4·5**	**2**	0·2	**0·3**	Li
Ni	8	0·07	—	0·07	**5**	—	—	0·07	**1·5**	—	—	0·07	**0·6**	3·6	14	2·1	20	**50**	**170**	150	**250**	Ni
Co	5	0·07	—	0·07	**5**	0·07	0·5	0·64	**13**	1·8	0·15	2·6	**22**	3·6	5·8	2·1	14	**35**	**53**	39	**60**	Co
Cu	20	3	1	4	**260**	3	15	22	**440**	21	—	43	**360**	16	8	21	88	**220**	**130**	42	**70**	Cu
Sc		0·02	—	0·02	**2**	—	0·4	0·42	**8**	0·6	0·1	1·1	**9**	1·8	2·4	0·7	6	**15**	**12**	6	**10**	Sc
Zr	350	8	0·15	8·2	**550**	0·25	0·5	8·95	**180**	0·9	1	10·8	**90**	2·7	4·0	7	24·5	**80**	**50**	25·5	**40**	Zr
Mn	1250	12	13	25	**1700**	37	40	102	**2000**	78	9	189	**1500**	95	84	120	488	**1220**	**800**	312	**500**	Mn
Y	250	2·6	1·5	4·1	**270**	0·7	—	4·8	**100**	—	0·6	5·4	**45**	—	—	—	5·4	**13**	*	—	—	Y
La	80	1·2	1·5	2·7	**180**	—	—	2·7	**55**	—	—	2·7	**23**	—	—	—	2·7	**7**	*	—	—	La
Sr	350	4·5	1·5	6·0	**400**	3	12·5	21·5	**430**	30	4·0	56	**470**	45	72	42	215	**500**	**350**	135	**220**	Sr
Ba	1150	8·5	1	9·5	**630**	0·75	1·3	11·5	**230**	2·7	6·0	20·2	**170**	4·5	2·2	9	36	**90**	**43**	7	**11**	Ba
Rb	150	0·8	—	0·8	**50**	0·15	—	0·95	**19**	—	0·2	1·15	**10**	—	—	—	1·15	**3**	*	—	—	Rb

Letters B, C, etc. to I and Y refer to average values given in Table F.
Dashes indicate small quantities which are neglected.
An asterisk in columns showing composition of 1st and 6th liquids indicates amounts less than sensitivity.
The actual weight of material in the various units of the intrusion would be obtained in metric tons by multiplying the figures, other than the percentages, by 10,000.

much as the original magma; V, Cu and Sr contents probably appreciably less, and a Ba content considerably less than the original magma. These differences are of the kind which would be anticipated from the previously established trends in mineral composition and the estimated major elements composition (1939, pp. 220–222).

Table 11—The estimated trace element composition in parts per million of the hidden layered series compared with other early rocks

	I *Av. chilled marginal gabbro* (*Y Table A*)	*II* *Av. early marginal border group* (*A' Table F*)	*III* *Estimated average composition of the hidden layered series* (*See Table* 10)	*IV* *Av. of earliest exposed layered rocks, the hypersthene-olivine gabbros* (*B Table F*)
P	280	220		260
Ga	17	15	**14**	19
Cr	170	300	**230**	230
V	140	170	**90**	220
Mo	*	*	*	*
Li	2	2	**0.3**	2
Ni	170	350	**250**	120
Co	53	80	**60**	48
Cu	130	50	**70**	67
Sc	12	10	**10**	20
Zr	50	30	**40**	33
Mn	800	(1100)	**500**	700
Y	*	*	*	*
La	*	*	*	*
Sr	350	300	**220**	600
Ba	43	15	**11**	18
Rb	*	*	*	*

The graph showing the trend in composition of the layered rocks and the acid differentiates (Figure 6) has been extended to include the estimated figures for the average hidden layered rock. Since the abscissae are proportional to the volume of the various rock types, the area between the curve for the trace element and the horizontal axis of reference is proportional to the amount of element in the whole intrusion and should equal the area of the rectangle whose sides are the horizontal line through the amount of the oxide in the original magma and the horizontal axis of reference. This is illustrated for vanadium in Figure 8. It is valuable to have these relationships in mind when using the graphs of the rock compositions (Figure 6).

The composition of the successive residual liquids has also been plotted against the percentages solidified (Figure 7). The curves obtained may be briefly discussed in groups showing similar tendencies. The falling curves such as those for Cr, Ni and Co are due to the amount precipitated in the early chrome spinel, pyroxene and olivine being greater than the amount present in the original magma. These curves resemble that for Mg among the major constituents (1939, Plate 27). A minor feature which was not anticipated, however, is the slight rise of the Cr and Ni curves for the latest liquids after having been virtually along the zero line for some distance. A second group of curves, those for Ba, Zr, Y, La and Rb rise slowly over much of the range and rapidly in the latest stages. Only small amounts of these elements are

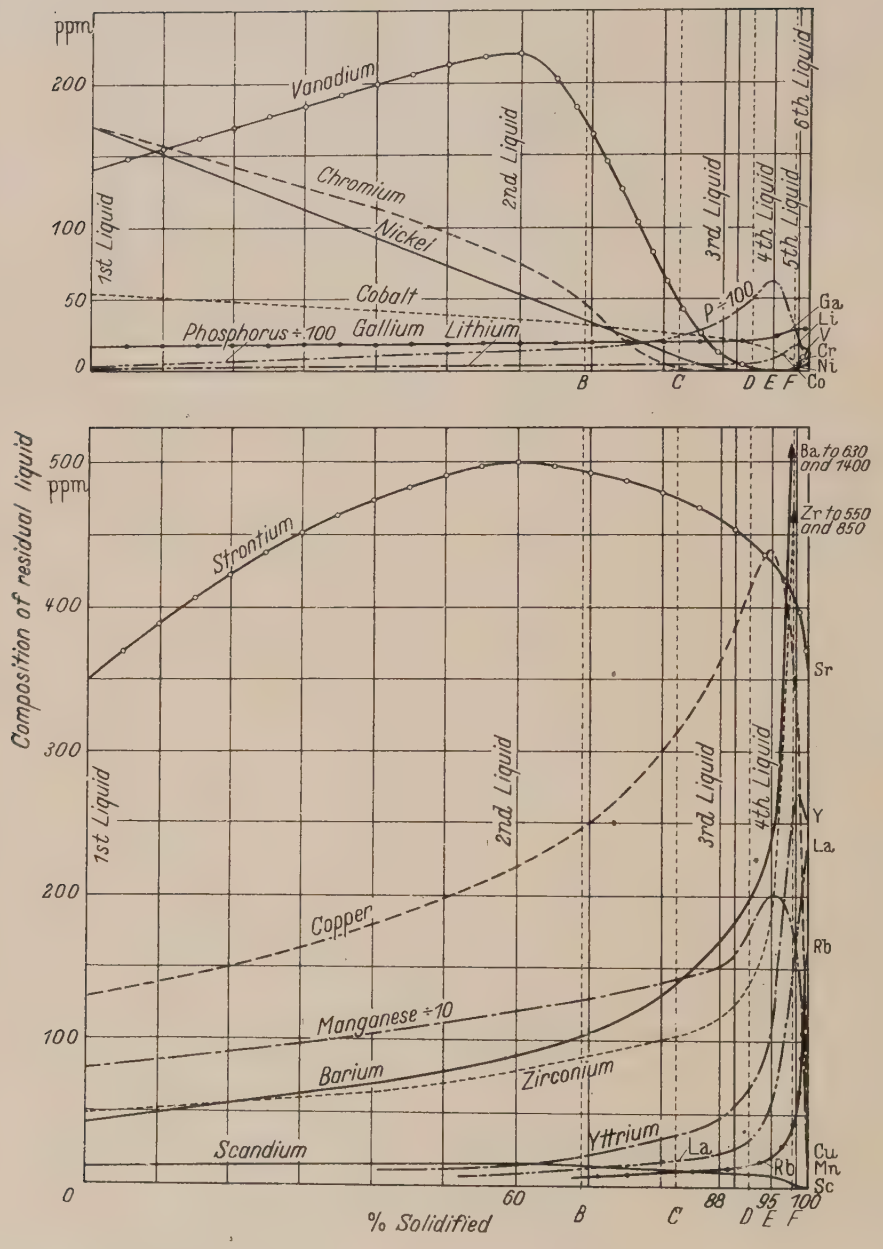

Fig. 7.—Variation in the estimated composition of successive residual magmas formed during differentiation of the Skaergaard intrusion. *B*, *C*, *D*, *E*, *F*, indicate the position on the graph of the following average rocks; *B:* hypersthene olivine gabbros; *C:* Middle Gabbro (without olivine); *D:* hortonolite ferrogabbro; *E:* ferro-hortonolite ferrogabbro; *F:* fayalite ferrogabbro.

precipitated in the early minerals and thus the amounts in the successive residual liquids rise. If no significant amount of an element is precipitated in the crystal fractions or if a constant small amount is precipitated, the resultant curve should have an approximately hyperbolic form, such as shown by these elements. Another group of curves, those for V, Sr and Sc, rise first and then fall. In these cases the amount of the element separated in the early minerals is less than the amount in the liquid, but in the middle stages the amount separated becomes greater than the amount in the residual magma and the curves fall. Cu, Mn and Li also behave in this

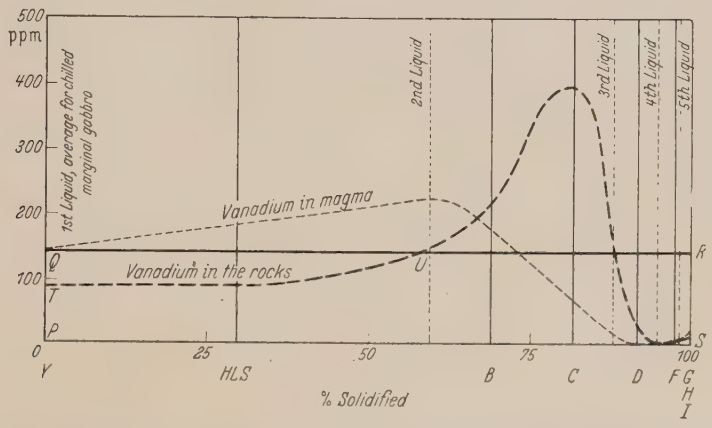

Fig. 8—Curves showing the variation in vanadium content of the Skaergaard rocks compared with the estimated vanadium content of the successive residual liquids.

way but the inflexion only occurs in the latest stages of the differentiation series. Among the major elements considered previously, Fe, Ti and P show this effect (1939, Plate 27).

Comparison of the curves for the successive liquids (Figure 7) and for the successive rocks (Figure 6) emphasizes the difference which must exist between the composition of rocks formed mainly by crystal accumulation and those produced by consolidation of magma without fractionation. Among the Skaergaard rocks the border group, except for the gabbro picrites and the perpendicular felspar rocks, are considered to be mainly the result of solidification of the liquid without crystal fractionation. The trace element compositions support this view. Thus the border rock 1837 (Table C) corresponds roughly with the magma at about 25% solidified. (Rock 4289 should be similar, but this is not the case with the amounts of Cr and V; as previously mentioned the sample analysed was probably too low in pyroxene to represent the material fairly.) The upper border rocks 3052 and 3050 resemble the composition of the second liquid when 60% had solidified, while the upper border rock 4163 resembles the third liquid when 88% has solidified, although in its low Cu and high Ba it resembles average grey gneiss (Table 9). Contamination with grey gneiss may well be the cause of the high Ba but cannot be the only factor involved in the low copper content.

X. The Sorting out of the Trace Elements during the Cooling of the Intrusion

Previous work on the intrusion has shown the processes involved in the differentiation (1939, pp. 262–289 and this paper, Section II) and it is now proposed to

consider how these processes, acting on the originally homogeneous magma, have produced the trace element distribution which is to be observed in the different minerals and rocks of the complex. In many cases the amount of a trace element in a particular rock can be related to the size and charge of the ions in relation to the crystal structure of the minerals making up the rock, and to the sequence of formation, relative amounts and major composition of the minerals. The full data on the trace elements are available in the tables, especially B to F, but it will often be sufficient to refer to the graphical presentation of the results (Figure 5 for the minerals, Figure 6 for the rocks and Figure 7 for the residual liquids).

No precise definition has been given or indeed seems to be possible for the terms major and trace constituents (*cf* Clarke and Washington, 1924, p. 5). In the case of minerals, the major constituents may be considered those essential to the crystal structure, or those contributing largely to it; the trace or minor constituents would then be any others which are present. If this definition be accepted then the same element may be a trace constituent in one mineral and a major constituent in another. In considering the rocks, on the other hand, the term trace element or trace constituent is used simply for an element present in relatively small amounts*. In general in this paper, what we have called the major constituents are those which have controlled the nature of the crystal phases separating from the magma; these elements have usually been determined chemically. What we have called the trace elements are those which are present in such small amounts that they have not usually affected significantly the sequence or amount of the crystal phases formed; these elements have usually been determined spectrographically. The distribution of the trace elements, as here defined, is largely determined by their relation to the major elements which control the various crystal phases formed from the magma. Phosphorus, chromium, zirconium and sulphur, however, have been included as trace elements although at certain stages they have formed small amounts of accessory minerals in which they are essential constituents.

In the previous paper on the Skaergaard intrusion the distribution of the chemically determined major elements, Si, Al, Ti, Fe^{+++}, Mg, Fe^{++}, Na, Ca and K has already been considered. Phosphorus and manganese were also determined by chemical methods and they too have been discussed. Sulphur and some other elements which were only determined chemically in special cases received no general consideration. Although we have no additional data for phosphorus and titanium the results will be recapitulated in order to make comparisons with the behaviour of the trace elements. For sulphur, which has proved of a special significance, some further data have been obtained by chemical means, and these allow generalizations on the distribution of sulphur throughout the intrusion to be made. For manganese, the data are largely chemical but as we have some additional spectrographic information manganese will be reconsidered. Having recapitulated the results for phosphorus and titanium, and having dealt with sulphur, we shall consider the other elements in order of ionic size at their usual state of oxidation in minerals.

1. Phosphorus

The amount of phosphorus precipitated with the early rocks (Table F and 1939, pp. 227–228 and Plate 27) was small, and less than the amount in the magma. Hence in the early and middle stages the phosphorus content of the magma rose. At a fairly late stage, apatite suddenly began to form as a primary precipitate

* In considering minerals it might be well to talk of major and minor constituents and keep the term trace constituent or trace element for use in connection with rocks.

Table 12—Sulphur and copper contents of Skaergaard rocks and minerals

(Figures in parts per million)

		Chilled marginal gabbro			*A*	*B*	*C*	*D*	*E*	*F*	*G*	*H*	*I*
		1825	1724	*Av.*									
Sulphur content of rocks		400	630	**500**		450	400		600*	4100	450		
Copper content of rocks	Spectroscopic determinations	80	150	**130**	45**	67**	180**	350**	600	300**	500**	350**	20**
	Colorometric determinations								860	330	570		
Copper content of minerals (spectrographically determined, see Table B)	Plagioclase				15	35	15	250	700	10			
	Pyroxene				< 10	35	100	300	1000	100			
	Olivine				20	20		300	400	20			
	Ilmenite					100***	30	300	400				
	Magnetite					20***	50	500	300	400****			

Sulphur and colorometric copper determinations by E. A. VINCENT. Spectrographic determinations by R. L. MITCHELL.
* Determination for rock 4144 and the others are for rock 4145.
** Averages for the particular stage.
*** Separated from the high hypersthene-olivine gabbro 2308.
**** May contain some pyrrhotite and associated copper minerals.

Column headed *A* = *Early border gabbro* 1851 *or* 1846
B = *Hypersthene olivine gabbro* 4077
C = *Middle gabbro* 3662
D = *Hortonolite ferrogabbro* 1907
E = *Ferro-hortonolite ferrogabbro* 4144 *or* 4145
F = *Fayalite ferrogabbro* 4142
G = *Basic hedenbergite granophyre* 4137
H = *Hedenbergite granophyre* (*Average*)
I = *Acid granophyre* (*Average*)

mineral. This happened when the estimated amount of phosphorus pentoxide in the magma had reached about 1·5%. At the level where apatite becomes a primary precipitate there is a sudden jump in the phosphorus pentoxide content of the layered series from about 350 ppm to 8700 ppm, and thereafter there is a gentle decline. The estimated amount of phosphorus pentoxide in the successive residual liquids also declines after the beginning of the formation of apatite as a primary precipitate, because the phosphorus pentoxide then being precipitated exceeded the amount present in the liquid. In the early layered rocks it is likely that phosphorus does not appreciably enter into the crystal structure of any of the primary precipitate minerals, and the small amount of phosphorus pentoxide shown by chemical analysis is considered to have been the phosphorus pentoxide in the 20% of magma that filled the interstices around the primary precipitate minerals. On crystallizing this interstitial liquid produced rare, small apatites, one of which has been figured in its characteristic situation in rock 4077 (1939, Figure 24, p. 91).

2. Titanium

It has been shown that titanium behaves similarly to phosphorus (1939, p. 228 and Plate 27). Thus the amount of titanium precipitated in the earlier rocks, although some entered the pyroxenes, was less than the amount in the magma, so that titanium accumulated in the residual liquid until it reached about 2·5% at a late hypersthene-olivine gabbro stage. At this stage, ilmenite suddenly became a primary precipitate mineral, and the amount of titanium in the layered rocks rises abruptly to a maximum and thereafter falls slowly. As in the case of phosphorus pentoxide, the amount of titanium dioxide present in the liquid declined after the appearance of ilmenite as a primary precipitate because more titanium was being extracted in the primary precipitate minerals than was present in the magma.

3. Sulphur

Sulphur has been determined chemically in several rocks by Mr. E. A. Vincent and we are grateful to him for this additional data. The results, which are scattered in various tables, are assembled in Table 12. The earlier determination by Deer (1939, Table XLI) are several times the values now given, a difference which remains unexplained. At one time it seemed possible that some of the lower results might be due to loss of sulphur in a powdered rock which had stood for a long time, but we have eliminated this possibility by preparing new samples of the powders. Dr. Deer has himself re-determined the sulphur content of the Middle Gabbro (3662) and agrees with the revised figures.

Two determinations of sulphur in the chilled marginal gabbro have been made (Table 12); the figures are not very similar but we shall take the mean, 0·05% as the sulphur content of the original Skaergaard magma. The average value given by Tröger for gabbros is 0·20%. General scrutiny of the available data seems to show that the content of sulphur is lower in basalts than in gabbros, perhaps due to losses in the volcanic pipe or at the time of extrusion but such an explanation cannot apply to the low values in the Skaergaard magma.

In the lower and middle stages of the layered series the amount of sulphur present—about 400 ppm—is slightly less than the amount in the original magma, but by the ferrohortonolite-ferrogabbro stage the value has risen to 600 ppm. At the next stage, that is in the fayalite ferrogabbros, the amount of sulphur increases

abruptly to 4100 ppm, but it falls again in the later basic hedenbergite granophyre to 400 ppm.

When the rocks of the layered series are examined microscopically, sulphides are seen only rarely and in small amount, except in the fayalite ferrogabbro. In this rock the textural relations of the pyritous minerals (Figure 9) suggest that they are the result of the solidification of droplets of an immiscible sulphide-rich liquid

Fig. 9—Photomicrograph of polished specimen of fayalite ferrogabbro 4142, showing composite sulphide mass of globular form embedded in silicates. Ragged masses of pyrrhotite (white, about 20%) with a little chalcopyrite (about 5%) are embedded in an aggregate of pyritous minerals, the nature of which at present is obscure (photo by R. PHILLIPS; magnification × 200).

formed within the silicate magma, rather than early crystal phases. Thus the bulk of the sulphides occur as rounded complex patches consisting of pyrrhotite, chalcopyrite, *etc* surrounded and veined by a fine-grained aggregate of pyritous minerals perhaps also containing silicates*. Further indirect evidence for the formation of an immiscible liquid sulphide phase is obtained from the variation in the copper content of the minerals and is discussed below (pp. 191).

The amount of sulphur present in the magma at the fayalite ferrogabbro stage when immiscible droplets of pyrrhotite-rich liquid apparently formed, can only be estimated roughly, but this is worthwhile doing as it is a figure of considerable interest. The number of sulphur determinations is not enough to allow figures for sulphur to be included in Table 10 in which the amount of various elements in successive residual

* For help with the preliminary investigation of the sulphides, we are indebted to Mr. R. PHILLIPS of the *Department of Geology*, Durham Colleges who is continuing work on these minerals.

liquids is presented, but the methods used in compiling the table can be applied to obtain an approximate value for the sulphur content of the liquid from which the fayalite ferrogabbros formed. Thus if the last $2\frac{1}{2}\%$ of the Skaergaard magma to solidify is taken as having a composition corresponding to 1% fayalite ferrogabbro, sulphur content 4100 ppm, and $1\frac{1}{2}\%$ of various later rocks with a sulphur content of 500 ppm or less, then the amount of sulphur in the last $2\frac{1}{2}\%$ of residual magma should be 2000 ppm or somewhat less. Apparently, therefore, liquid immiscibility occurred when the amount of sulphur in the magma was about 0·2%. Vogt (1921, p. 644) summarizing the results of previous experiments and thin section investigations, concluded that, if the amount of ferrous sulphide in gabbroic magma at a temperature of 1350°–1400° C was greater than about 0·25 to 0·4% (corresponding to a sulphur content of 0·08 to 0·13%), then a pyrrhotite-rich liquid would form as a distinct liquid phase, and that at lower temperatures the amount to cause a separate sulphide liquid to form became rapidly less. In the fayalite ferrogabbro of the Skaergaard intrusion, only a proportion of the sulphide is pyrrhotite, but the indication is that immiscibility of the sulphides in the silicate magma occurred when the amount of sulphur was of the same order as given by Vogt.

The lower and middle layered rocks have a sulphur content rather less than that of the original magma, and thus with fractionation the amount of sulphur increased in the successive residual liquids. Microscopic investigation of the lower and middle layered rocks has so far not revealed any sulphur minerals except occasional fragments of bornite and the situation of the 400 ppm of sulphur in these rocks is not at present known. As a result of the postulated increase in the amount of sulphur in the magma, a stage is ultimately reached when droplets of an immiscible sulphide liquid are believed to have formed. This was at a late stage when probably $97\frac{1}{2}\%$ of the Skaergaard magma had solidified and the sulphur content of the residual magma was about 0·2%. Dense sulphide droplets, forming in this late residual liquid, would tend to sink and they were apparently precipitated along with the crystal phases which formed the fayalite ferrogabbro. The removal of sulphide as immiscible liquid droplets would reduce the sulphur content of the later residual liquids and therefore also of the later rocks. The determination of sulphur in one of the later rocks, the basic hedenbergite granophyre, gave a value of 500 ppm, which is in general agreement with this postulated sequence of events. In layered intrusions the formation of an immiscible liquid sulphide phase has been suggested before, for instance, by Wagner for the Bushveld complex (1929, pp. 107, 109, 239) and by Scholtz for the Insizwa and related masses (1936, pp. 204–205).

4. Gallium

Gallium enters plagioclase and magnetite fairly readily and the estimated amount in the successive residual liquids rises only slowly. The high content in the latest plagioclase is probably due to the larger amount in the liquid combined with increasing ease of entry as indicated by the distribution factor 7. In the plagioclase the Ga ion is replacing the smaller Al ion and an increase of the amount in the late stage would be expected. The lack of any increase in the amount of gallium in the late magnetites is no doubt due to the decrease in the ease of entry in the latest stage as indicated by the distribution factor 1. In magnetite the Ga ion is replacing the somewhat larger Fe^{+++} ion and this contrasted behaviour of gallium in plagioclase and magnetite is as would be expected.

Since plagioclase is much more abundant in the rocks than magnetite, it is the amount of gallium in the plagioclase which is mainly responsible for the gallium

content of the rocks. Ga largely replaces Al in the rock-forming minerals and therefore GOLDSCHMIDT and PETERS (1931, and *cf* GOLDSCHMIDT, 1934) estimated the average atomic ratio Ga : Al for various rocks; for gabbro the average ratio is 1 : 40,000 and for granite it is 1 : 16,000. The ratio for the gabbroic rocks of the Skaergaard intrusion is 1 : 15 000 and it increases with differentiation until in the acid granophyres it is 1 : 6,000. For the various Caledonian plutonic rocks, largely granites, the ratio averages 1 : 12,000 (NOCKOLDS and MITCHELL, 1948). The curves for Ga and Al (Figure 6) show a considerable degree of sympathetic variation but they also indicate a slight increase in the ratio of Ga : Al as differentiation proceeds.

The average given by GOLDSCHMIDT and PETERS for the amount of gallium oxide in gabbros is 0·001%. The average amount of gallium in the original Skaergaard magma is 17 ppm which corresponds to 0·002% gallium oxide.

5. Chromium

This element is strongly concentrated in the early rocks: the gabbro picrites and the eucritic gabbros of the outer border group; it is virtually absent in the later middle stages, the middle gabbros and ferrogabbros, but there is a return to small amounts in the latest differentiates, the granophyres.

Although the chromium content of the original magma was only 170 ppm corresponding to 0·025% chromic oxide, an opaque chrome-spinel separated as a distinct phase, for a brief early stage in the differentiation process. The chrome-spinel is concentrated along with olivine in the gabbro-picrite, an early rock of the northern border group which has 1500 ppm of chromium (0·23% Cr_2O_3). VOGT (1921, p. 322) pointed out the extreme insolubility of chromite in silicate magmas. Precipitation of chromite from a magma with only 0·025% chromic oxide would seem to be about the lower limit for the separation of chrome-spinels, since in the Skaergaard sequence chromite soon ceases to be a separate phase; for instance, it does not seem to be present in such early marginal border rocks as 1851 and 1846. VOGT (1921, p. 323) postulated that chromite would form as a separate phase if only 0·1 or 0·05% Cr_2O_3 were present in the magma. Our figure of 0·025% chromic oxide reduces still further the lower limit of chromium necessary for the separation of the chrome-spinel. The low solubility of chromite in gabbroic magma is not very different from that of zircon in granite magma.

Besides forming early chrome-spinel chromium enters abundantly into the early pyroxenes, 3000 ppm being present in the earliest pyroxene analysed, and 350 ppm being still present in the pyroxenes of the hypersthene-olivine gabbro stage. In olivines separating simultaneously with these chromian pyroxenes the amounts of Cr present are 20 ppm and < 1 ppm respectively. It is not certain whether the 20 ppm chromium in the earliest olivine is present in the crystal structure or is due to contamination by pyroxene (see pp. 152–153), but in any case the emphatic preference of chromium for pyroxene rather than olivine is clear.

Magnetite and ilmenite are not formed as primary precipitate minerals until late in the hypersthene-olivine gabbro stage, by which time the amount of chromium in the magma is small. At this stage what chromium there is in the magma enters the magnetite, giving crystals with 200 ppm chromium, while the simultaneously formed ilmenite contains only 2 ppm chromium and some of this may be due to contamination. Ilmenite and magnetite from the Middle Gabbros contain less than 1 ppm chromium. At about the stage when magnetite becomes a primary precipitate it is estimated that the chromium content of the magma is only a few parts per

million (Table 10 and Figure 7). Thus in the case of the Skaergaard intrusion, after early precipitation of chrome-spinel, there is a break when no spinel mineral formed as a primary precipitate. Then at a relatively late stage the iron spinel, magnetite, began to precipitate and this received most of the chromium that was still available in the magma. This sequence of events may be similar to the conditions in the Bushveld complex where chromite bands occur fairly low in the critical zone and magnetite bands considerably higher up. However, there is a significant difference in that the Bushveld magnetite has a chromium content of 2000–3000 ppm while the Skaergaard magnetites are chromium-poor (pp. 158–159).

The crystallization break between the early chrome-spinels and the later ferric iron spinels is interesting in view of the incomplete miscibility for these two spinels postulated by the editors of the new edition of Dana (1944, p. 688). The supposed immiscibility, which does not seem to be indicated by the work of Stevens (1944, pp. 31–33), may be due to the fact that conditions in igneous rocks usually give a break in the period of crystallization, and also a break in the composition which, is not, however, the result of immiscibility.

Chromium enters abundantly the early spinelid mineral, and the early pyroxene, and this is in general agreement with the smaller size of Cr^{+++} (0·64Å) compared with Fe^{+++} (0·67Å). The chrome-spinel and chromian pyroxene separated in sufficient quantities to cause a rapid reduction of the Cr content of the magma and chrome-spinel soon ceased forming. Although the tendency of chromium to enter pyroxene and magnetite, as judged by the distribution factors, may well remain high, yet later pyroxene and iron ores contain only very small amounts because there is so little in the liquid from which they are forming. The low content of chromium in the late pyroxenes and iron ores is the effect of fractional crystallization and is not necessarily due to any difficulty of entry of Cr^{+++} into the lower temperature pyroxene and spinel solid solutions.

The completeness with which chromium is removed from the magma is surprising, the amounts being estimated to be reduced at certain stages well below 1 ppm. Even if some of the data used in constructing Table 10 be open to doubt, it is clear that the amount in the residual liquids at certain stages must be very small, since analyses of several rocks of the layered series which contained about 20% of interstitial magma have less than 1 ppm of chromium over the range from hortonolite ferrogabbro to fayalite ferrogabbro (Table C). This indicates that the 20% of interstitial magma has less than 1 ppm of Cr and that the magma had certainly less than 5 ppm chromium. In the latest granophyre differentiates, however, an average amount of 20 ppm chromium is again found. It appears that late pyroxenes, olivines and iron ores removed so little chromium that it began to accumulate again in the residual liquids. Such an effect might be due to the distribution factor for chromium between the late iron-rich crystals and the liquid being significantly less than that for the magnesium-rich minerals but the reasons for such a change would be difficult to understand. Another possibility is that in these liquids the chromium is in a different state of oxidation* and although Cr^{+++} enters readily, more oxidized chromium does not. If the hypothesis of the chromium being in a more oxidized state is right, then the distribution factors calculated on the assumption that the chromium in the liquid is all Cr^{+++}, will not be correct. This may account for the trend to low distribution factors indicated for chromium in

* Vanadate and chromate minerals of mineral veins suggest a change to a higher state of oxidation for chromium and vanadium at some stage in the differentiation process.

the pyroxenes (Table 6) but rather similar phenomena are shown by nickel, where an hypothesis of a change in the state of oxidation is not feasible and the problem must unfortunately be left in an unsatisfactory state.

The location of the chromium in the granophyres is not known. It may be present in the stilpnomelane or the chlorites but these have not been analysed.

The average chromium content of gabbro is given by GOLDSCHMIDT (1937, p. 662) as 0·05% chromium oxide. The amount of chromium in the original Skaergaard magma is rather less, being 170 ppm corresponding to 0·025% chromium oxide. The average chromium content of acid rocks according to GOLDSCHMIDT is 3 ppm chromium oxide (1937, p. 662). A figure of 20 ppm chromium oxide was, however, obtained by TONGEREN (1938) for granites from the Dutch East Indies and the values for various granites given by RANKAMA (1946) and NOCKOLDS and MITCHELL (1948) show that many have significantly greater amounts than 3 ppm. Thus the figure of 20 ppm for the late granophyre differentiates of the Skaergaard intrusion, although greater than GOLDSCHMIDT's average, is paralleled by those of many other granite rocks.

6. Vanadium

Vanadium, like chromium, is a characteristic constituent of the pyroxenes and iron ores, and not of the olivines; it is probably also present in small amounts in plagioclase. Of all the minerals of the gabbros, vanadium has a decided preference for entering magnetite; thus in the late hypersthene-olivine gabbros where magnetite, ilmenite, pyroxene and olivine are all separating as primary precipitate minerals the parts per million of vanadium in these minerals are respectively 2000, 600, 200 and < 5. After this stage the amount of vanadium in the various series of minerals falls off, although not so rapidly as chromium; late pyroxenes, olivines, ilmenites and magnetites have less than 5 ppm of vanadium. It would be expected that V^{+++} having an ionic radius of 0·65 compared with 0·67 for Fe^{+++} would be received readily by early minerals containing ferric iron but that the tendency to concentrate in the early minerals would not be so considerable as that of Cr^{+++} of ionic radius 0·64. In this respect the behaviour of these elements in the Skaergaard differentiation series is as anticipated. The absence of vanadium in the late pyroxene and iron ores is, no doubt, not due to any lessening of the tendency to enter these crystals, but is the result of the reduced vanadium content of the magma from which they were forming. The rather small removal of vanadium from the magma in the early stages results in an increase in the vanadium content of the magma and this maintains the vanadium content of the pyroxene at a high value. However, formation of magnetite with a high content of vanadium, as a primary precipitate in the middle stages of the fractionation, causes a fairly rapid decrease in the amount of vanadium in the magma, until it is so low that the vanadium content of the pyroxene and iron ores falls to values below the sensitivity, 5 ppm.

The conditions in the Skaergaard intrusion should be favourable for the concentration of vanadium in primary magnetite because the late formation of magnetite allowed a rise in vanadium content of the magma before magnetite formed; yet the content of vanadium trioxide in the Skaergaard magnetite only reaches a maximum of 0·4%. This is approximately the amount found by HUTTON for certain titaniferous magnetites of New Zealand (1945, p. 298). On the other hand, vanadium-rich, magnetite, iron ores from India examined by DUNN and DEY (1937) had a V_2O_3 content reaching 7%. FRANKEL and GRANGER (1940, p. 105) have suggested

that the Indian, vanadium-rich iron ores are due to the presence of vanadian maghemite. Whether this is so or not, the Skaergaard evidence suggests that vanadium-rich ores are not the result of simple magmatic segregation but of special processes as suggested by Dunn and Dey.

The amount of vanadium in the original magma is 140 ppm. There is no recent estimate of the vanadium content of gabbros but an average figure of 0·04 V_2O_3 (=250 ppm V) is given by Tröger (1935). The amount of vanadium separating in the early pyroxenes is 300 ppm and in the other early minerals it is only 10 ppm or less. Since the amount of pyroxene in the early rocks is only about 25%, the amount of vanadium precipitated in the early rocks is less than the amount in the magma and the vanadium content of the successive residual magmas increases until it reaches an estimated value of 230 ppm (Figure 8). When iron ores become a primary precipitate, late in the hypersthene-olivine gabbro stage, much vanadium is extracted in the iron ores and the average layered rocks then contain 400 ppm of vanadium. From this point onward the amount of vanadium in the magma falls and soon reaches low values, estimated at less than 5 ppm. As with chromium, there is an increase in the amount of vanadium in the latest differentiates, reaching about 20 ppm in the granophyres. To account for this we can only put forward the same tentative hypotheses as made for chromium.

In a less reduced magma than the Skaergaard, the iron ores would probably separate earlier and there would not be a concentration of vanadium in the residual magma of the middle stages. For this reason fractionation of magma having a low state of oxidation like that of the Skaergaard should provide especially good conditions for the development of primary magnetite rock of a high vanadium content; even so the indications are that the amount only reaches 0·3% V_2O_3.

7. *Molybdenum*

The amount of molybdenum in the rocks of the Skaergaard intrusion is about the sensitivity (1 ppm) or below over nearly the whole range. The figure of 3 ppm for the gabbro picrite at one end of the differentiation series and the similar figure for certain of the granophyres at the other end are, however, regarded as significant. Some concentration of molybdenum in late differentiates is to be expected since molybdenite is a common constituent of granite pegmatites. Rather surprisingly the mineral analyses show the presence of appreciably greater amounts of molybdenum in olivine than in contemporaneous pyroxenes or iron ores, and this is probably the reason for the higher amount of molybdenum in the gabbro-picrites. It is interesting that Hawaiian olivine basalts investigated by Fergusson (1914) also contain relatively high molybdenum (0·01% MoO_3). From a number of observations on rocks with a wide range of silica percentage, Sandell and Goldich (1943, p. 168) showed that the molybdenum content is about 1 ppm over most of the range they investigated but becomes considerably more in the rocks with more than 75% SiO_2.

It is not easy to offer any hypothesis to account for enrichment of molybdenum in the olivines which would not equally apply to the pyroxenes. On the whole, tri- and quadrivalent elements of appropriate size enter the pyroxenes, where valency adjustments are easily made, rather than the olivine. Thus we expected molybdenum in the pyroxenes and not in the olivines. However, there can be little doubt that molybdenum is more abundant in the olivines than in the pyroxenes and we tentatively suggest that it is present replacing silicon within the oxygen tetrahedra.

It may be that this replacement is more easily made in olivine because its isolated SiO_4 groups are more capable of slight expansion, than the SiO_4 groups in the pyroxene chains. Wickmann (1943, pp. 375–378) has made a similar suggestion to account for some of the features shown by the distribution of germanium.

8. Lithium

Lithium remains fairly constant at the low value of about 2 ppm in all the early and middle differentiates but in the acid differentiates, it rises to 6 ppm and finally to 20 ppm. Strock (1936) gives rather different average values: 10 ppm for average gabbro and basalt and 150 ppm for average granite.

In the early and middle stages of the differentiation lithium is present in about the same amount in plagioclase, pyroxene and olivine, and this has been taken as an indication that the lithium content is a result of some special kind of contamination perhaps connected with slight hydrothermal alteration. The mineral analyses show lithium in definitely increased amount in the latest pyroxenes and olivines and this conforms with Strock's observations (1936) that lithium replaces magnesium and iron in late stage ferro-magnesian minerals, the late entry being the result of the low charge on the Li ion. The ratio $Li_2O/MgO \times 10{,}000$ for certain stages of the differentiation of the Skaergaard rocks is as follows: early border gabbro 0·4, hypersthene-olivine gabbro 0·4, hortonolite ferrogabbro 0·8, fayalite ferrogabbro 50, acid granophyre 170. These figures may be compared with those for average rocks given by Strock (1936, p. 197): gabbro 3·3, diorite 10, granite 600.

Whatever the manner of occurrence of the lithium in the minerals, the amount extracted in the earlier rocks is smaller than the amount in the original magma and lithium becomes concentrated in successive residual magmas up to the penultimate stage, after which there is a slight fall (Figure 7).

9. Nickel

In a prescient paper, Vogt (1923) considered the occurrence of nickel in igneous rocks and showed that: (1) nickel normally occurs in magnesium silicates (2) the amount of nickel tends to increase with increasing magnesium content and to decrease with increasing iron content and (3) in silicates having the same ratio of Mg/Fe, the olivines contained more nickel than the ortho-pyroxenes, and the clino-pyroxenes and amphiboles successively less. Our results, so far as they go, are in agreement with Vogt's findings.

The ease of entry of nickel into pyroxene as judged by the distribution factors, is about twice that of magnesium, and the ease of entry of nickel into olivine while being different from that for pyroxene, is again about twice that for magnesium. In comparing the amounts of nickel in the ferro-magnesian minerals and iron ores, it is clear that, at comparable stages, the amounts in the pyroxene are about half those in olivine and magnetite. For ilmenite, the values seem to be about the same as for pyroxene, but these data may not be satisfactory. The reason for the smaller amount in pyroxene is no doubt the replacement by nickel in the Mg/Fe positions, but not in the Ca positions; thus there is only about half the opportunity for nickel to enter pyroxene as olivine or magnetite. Whether there is any real significance in the apparent differences in the amount of nickel in ilmenite and magnetite is doubtful, especially as the amounts of cobalt are about the same in olivine, magnetite and ilmenite. It seems likely that our figures for nickel in the ilmenites are rather low.

The distribution factors apparently increase in all four mineral series with fractionation and as considerable quantities of these four minerals are precipitated, the

nickel content of the magma is gradually reduced to a very low value. The low amounts of nickel (less than 2 ppm) in late pyroxenes, olivines, ilmenites and magnetites are not due to any difficulty of entry but to the liquid having become depleted in nickel by the abundant entry of this element into earlier fractions. The granophyres, however, show a return to a small nickel content of 5 to 10 ppm. The completeness with which the nickel is eliminated from the magma by its entry into the magnesium-iron minerals is paralleled by chromium. In the Skaergaard rocks nickel is reduced to less than 2 ppm while the magnesium oxide content is still 5·5%. Almost complete elimination of nickel in what are taken as late differentiates of basic magma has also been shown by Lundegårdh (1946 and 1949, pp. 9–16). In the case of Cr^{+++} early entry is anticipated from the ionic size and charge. For nickel the ionic size in sixfold coordination is usually given as 0·78Å and this is the same as that for Mg^{++}. Goldschmidt, however, has pointed out (1944), that nickel has a tendency to homopolar bonding which increases the effective bond strength* compared with magnesium and he gives this as the reason for the preferential entry of nickel.

By means of the distribution factors (Table 6) we have shown the degree of preferential entry of nickel compared with certain major elements. In studying mineral suites such ratios will often be unobtainable, and then it is of value to give direct ratios between the trace element and some related major element which is present in the mineral. As an example the ratio $Ni/Mg \times 1000$ for the Skaergaard pyroxenes and olivines is given in Table 13 and the values may be usefully compared with data for Caledonian igneous minerals given by Nockolds and Mitchell (1948). Under the conditions of strong fractionation which occur in the Skaergaard intrusion this ratio for the pyroxenes varies from 0·03 to 2·1, that is, by a factor of about 100, and for olivines it varies from 0·04 to 9·5, a factor of 200. Some series of minerals separated from related rocks show very much less variation. Thus the ratio for pyroxenes, hornblendes and biotites from Caledonian plutonic rocks shows only a variation from 0·6 to 1·8 (Nockolds and Mitchell, 1948, Tables XII and XV). If in a mineral or rock series, this ratio shows little variation, then there has probably been no appreciable fractionation, while much variation, as in the Skaergaard intrusion, indicates strong fractionation.

In Table 13 we have presented other ratios between trace elements and major elements for the Skaergaard pyroxenes and olivines to show the way such ratios vary under conditions of strong fractionation and to allow easy comparison with the data of Nockolds and Mitchell (1948) and Lundegårdh (1946, Figures 45–50 and 1949).

10. Cobalt

In the original magma this element is only about one third as abundant as nickel. This is roughly the ratio given by Goldschmidt for the relative abundance of nickel and cobalt in igneous rocks, namely 100: 40. In an early estimate, Clarke and Washington (1924) gave this ratio as 20: 1 and later Vogt (1931) gave it as 10: 1. The cobalt content of average gabbro is estimated by Goldschmidt as 0·01% CoO; the amount in the original Skaergaard magma is near this, being 50 ppm of Co, *i.e.* 0·007% CoO.

* As evidence of the bond strengths Goldschmidt (1944, p. 4) quotes the following cell sizes:

Ni_2SiO_4	a 4·71	b 10·11	c 5·91 Å
Mg_2SiO_4	a 4·77	b 10·26	c 5·99 Å
Fe_2SiO_4	a 4·80	b 10·59	c 6·16 Å

Table 13—Ratios of trace, to major elements in pyroxenes and olivines
(For actual amounts see Table B)

	Pyroxenes						*Olivines*					
Stage of differentiation (see p. 136)	*A*	*B*	*C*	*D*	*E*	*F*	*A*	*B*	*C*	*D*	*E*	*F*
Cr× 1000/Fe +++	220	220	>0·2									
V× 1000/Fe+++	22	160	20	3	<0·6							
Cr× 1000/Mg+++	31	4	0·02					<0·005				
V× 1000/Mg	3	2·8	1·4	0·5	<0·2			<0·03				
Ni× 1000/Mg	2·1	1·6	0·7	<·03	<0·03		9·5	1·8		0·09	<0·04	
Co× 1000/Mg	0·63	0·56	0·83	0·62	1·1	4·7	0·72	0·67		0·92	1·0	3·3
Zr× 1000/Mg	0·3	0·6	0·4	0·8	<0·3							
Co× 1000/Fe++	0·9	0·5	0·4	0·29	0·16	0·06	0·75	0·5		0·27	0·11	0·04
Cu× 1000/Fe++	<0·15	0·35	0·7	2·1	5·5	0·4	0·1	0·08		0·8	0·9	0·04
Sc× 1000/Fe++	1·2	0·3	0·3	1·1	0·4	0·3	<0·05	<0·04				
Mn× 1000/Fe++	22	21	19	13		25	10	7			14	15

Cobalt enters the same minerals as nickel. As with nickel the amount of cobalt in olivine is about twice as much as in pyroxene separated simultaneously from the magma. Ilmenite and magnetite show amounts of cobalt about equal to that in simultaneously precipitated olivine.

Cobalt remains much more constant in amount in the various minerals than nickel. The distribution factors in the early fractions are similar to those for Fe^{++}. The small but steady decrease in cobalt in the ferro-magnesian minerals and iron ores during fractionation results from reduction in the amounts of cobalt in the successive residual liquids. The amount of cobalt removed by the pyroxene, olivine and iron ores—minerals which are precipitated abundantly—is sufficient to cause a decrease in the cobalt in successive residual liquids and despite the slight rise in the distribution factors with fractionation, there is a decline in the amounts of cobalt actually present in the minerals.

Table 14—Ratio of nickel to cobalt in Skaergaard rocks and in the successive residual magmas

Stage of differentiation (see page 136)	*A*	*B*	*C*	*D*	*E*	*F*	*G*	*H*	*I*
Ni/Co in the rocks	11	2·5	1	0·03	0·05	0·1	0·3	2	2
Ni/Co in the residual magmas (estimated)	3	1·5	0·03		0·1			2	

The amount of cobalt in the layered rocks falls only slightly during fractionation compared with the rapid fall of nickel, and the estimated cobalt content of the successive residual liquids also falls slowly compared with nickel. For the layered rocks, representing mainly successive crystal fractions, and for the successive residual liquids the Ni/Co ratio changes markedly during differentiation (Table 14). As would be expected the change in this ratio for the rocks is more marked than for the liquids. The significance of the Ni/Co ratio has been stressed by LUNDEGÅRDH, and he suggests that it can be used to estimate the stage reached in a differentiation series (1946, pp. 150–151). Our data confirm that the ratio should be of use in the basic and intermediate rocks, but in acid rocks where the actual amounts of these elements

are small, the ratio may not be significant. The increase in the ratio shown by the Skaergaard data towards the acid end of the differentiation series is due to the return of slight amounts of nickel and not to any change in trend in the amount of the cobalt. The reason for the slight increase in the amount of nickel in the acid rocks is not understood. For the acid rocks the ratios shown by the Skaergaard intrusion correspond well with those given by Goldschmidt for average granite (3: 1) and also with the ratio noted by Lundegårdh. The absolute amount of cobalt in the granitic differentiates of the Skaergaard intrusion averages 0·0006%, which again corresponds well with Goldschmidt's average of 0·001%.

11. Copper

In the original magma 130 ppm of copper is present. The range of values for copper in basic rocks indicated by Sandell and Goldich (1943) is considerable, but from their graphs a rough mean may be taken as 150 ppm. Tröger (1935) gives 0·03% CuO as the average for gabbros. Apparently the copper content of the Skaergaard magma is about average for basic rocks.

The amount of copper in the early rock fractions is definitely less than the amount in the magma, resulting in an increase in the copper content of successive residual magmas until an estimated value of 450 ppm of copper is reached at a stage when 95% of the magma has solidified (Figure 7). After this the amount falls and in the latest fraction, the acid granophyre, the amount is only 20 ppm. In this connection it is interesting that Sandell and Goldich's estimate of the copper content of twenty different acid rocks is 16 ppm.

The copper content of the various minerals (Table 12) supports the hypothesis of an immiscible liquid sulphide phase already discussed (pp. 180–182); indeed, it was consideration of the variation in the amounts of copper in the various silicate minerals which led to the discovery of the presence of the abundant sulphides at the fayalite ferrogabbro stage. In the early differentiates, including the lowest accessible hypersthene olivine gabbro of the layered series, the copper content of the plagioclase, pyroxene and olivine lies between 15 and 35 ppm and the distribution factors are round about 0·1. The copper is believed to be replacing sodium in the plagioclases and Fe^{++} in ferro-magnesian minerals and iron ores. For these same minerals, and also for the iron ores when they become primary precipitates, the copper content rises during fractionation until at the ferrohortonolite ferrogabbro stage (E) it lies between 300 and 1000 ppm and the distribution factors are about 1. In the subsequent fayalite ferrogabbro stage (F), the copper content of the various minerals drops to a tenth of the previous values (except for our sample of magnetite which may be impure, p. 156); at the same time the copper content of the rock remains fairly high. Our explanation of these facts is that at the fayalite ferrogabbro stage an immiscible liquid sulphide phase separated, into which the copper entered, thus greatly reducing the amount in the silicate liquid. On crystallization of the silicate liquid, the various minerals had necessarily only low copper contents because there was only a small amount of copper in the silicate liquid from which they formed. The minerals crystallizing from the sulphide liquid on the other hand would be expected to include some copper mineral and chalcopyrite has been identified. Ramdohr (1940, p. 18) states that the proportion of chalcopyrite in sulphide droplets due, it is believed, to immiscibility, lies between $\frac{1}{5}$ and $\frac{1}{8}$ and this may well prove to be about the proportion in the Skaergaard droplets when the nature of all the various minerals in them has been established.

On the hypothesis of the formation of an immiscible liquid sulphide phase, there should be a significant relationship between the distribution of copper and sulphur in the rocks. The high sulphur content of the fayalite ferrogabbro 4142 would be expected to correspond with a high copper content. Actually in the fayalite ferrogabbro, 4142, the amount of copper is only half that in the previous stage while there is eight times as much sulphur. This lack of direct correlation was so unexpected that we thought the spectrographic data might be at fault, especially as it seemed possible that during spectrographic analysis the copper present in the relatively volatile sulphides might have behaved differently from that in the refractory silicates. Colorimetric determinations of copper in certain of the rocks were therefore made, as shown in Table 12. The values fully confirm the spectrographic data. The copper content of the sulphur-rich fayalite ferrogabbro 4142 must be largely in the sulphides, but the total amount in the sulphides and silicates together does not come up to the value found for the immediately preceding ferrohortonolite ferrogabbro, where the copper is essentially only present in the silicates.

From the evidence of the copper content of the analysed silicate minerals of the fayalite ferrogabbro 4142, it seems clear that the silicate liquid from which they separated must have had a reduced copper content; we suggest that this reduction resulted from the formation of an immiscible liquid into which the copper entered abundantly so as to reduce the copper content of the silicate liquid. Unfortunately, among our collection of layered rocks there is none with a sufficient amount of copper sulphide in it to explain the postulated extensive removal of copper from the silicate liquid, but perhaps more detailed collecting just below the horizon of the fayalite ferrogabbro 4142 will reveal sulphide-rich fayalite ferrogabbros with high copper content.

The next differentiate after the fayalite ferrogabbro, the basic hedenbergite granophyre, has a relatively high copper content and only a small amount of sulphur (it is only in the acid granophyres that the copper content falls to the low value of 20 ppm). The high value of 500 ppm in the basic granophyres is difficult to explain. Sulphides have not been seen in these rocks (as is the case also with the greater part of the layered series which have only 450 ppm of sulphur) and the location of copper in these rocks is not at present known. The basic hedenbergite granophyre is considered to be essentially the result of the solidification of a residual silicate liquid and its high copper content suggests therefore that the late residual magma again developed a high copper content after its supposedly low content at the fayalite ferrogabbro stage. Some increase in the copper content of this residual magma might be due to further crystal fractionation allowing building up of the copper content of the liquid, but this process cannot be expected to account for the figure of 500 ppm. Perhaps processes other than crystal fractionation have been involved and it is clear that further collecting and analysis of these late stage rocks is required.

In some of the Skaergaard basic rocks the copper has been shown to be present in the silicate minerals, and in others it is mainly present as sulphide. Broderick and Hohl (1935, p. 304) showed that in the basalts of the Keweenawen the average amount of copper is 0·01% and of sulphide 0·013%, and they suggested the copper is largely present as sulphide. Newhouse (1936, p. 17) also concluded that in gabbro and basalt there is usually enough chalcopyrite visible to account for all, or nearly all the copper. On the other hand Reed (1936, p. 168) in a study of a basic rock from Alaska gave evidence that the location of most of the copper is in the silicate minerals and Ramadohr (1940, p. 36) gives it as his opinion that the copper of the

basic rocks occurs in the silicates as well as in the sulphides. The data given here for the Skaergaard intrusion, although not yet as satisfactory as they might be, seem to provide a clue to the conditions controlling the variable behaviour of copper in basic rocks.

12. Scandium

Goldschmidt and Peters (1931b) have shown that scandium is characteristically present in the pyroxene of igneous rocks and the data from the Skaergaard intrusion provides additional evidence. The scandium must be occupying the Mg/Fe or the Ca position in the structure. Scandium is also present in the late apatite. It is interesting how completely it is lacking in the olivines; this may be due to the difficulty of valency adjustment in the replacement of Mg and Fe by trivalent scandium or it may be due to the Sc preferentially occupying the Ca position. The Skaergaard data do not show any regular variation in the Sc content of the pyroxenes except perhaps a slight fall in the later fractions.

The content of scandium in the original Skaergaard magma is 12 ppm, equivalent to 0·002% Sc_2O_3. Goldschmidt gave 0·003% as the average amount of scandium oxide in gabbros. The scandium content of the rocks and of the residual magmas rise slightly with fractionation until a late stage when it slowly falls reaching figures below the sensitivity of our determinations (10 ppm) in the acid differentiates. The average value for scandium oxide in granites given by Goldschmidt and Peters (1931) is 2 ppm.

13. Zirconium

This element is present in the original magma to the extent of 50 ppm. No average data for zirconium content of gabbros seems to be available. Zirconium enters the early pyroxene in amounts varying from about 30 to 50 ppm, but it is absent in the later. At the hortonolite ferrogabbro stage, when apatite becomes a primary phase, zirconium enters the apatite in an amount about equal to that in the pyroxenes. It is still present in the later apatite but it is below the sensitivity (10 ppm) in the pyroxenes. There must be a very strong decline in the ease of entry of zirconium into pyroxenes because the amount present decreases strongly despite a rise in zirconium content of the liquid. The figures for the distribution factor of the pyroxenes show this satisfactorily. Zirconium apparently replaces calcium in early pyroxenes and apatites but not calcium in the plagioclases. The zirconium recorded as present in the latest plagioclase is thought to be due to contamination by zircon.

The amount of zirconium in the early rocks is about 30 ppm and this must be present partly in the pyroxene and partly in the inter-precipitate liquid which later solidified round the primary crystals. Since the amount of zirconium separating out in the early rocks is less than the amount in the magma, the zirconium content of successive residual liquids rises, reaching 150 ppm at about 95% solidified. Soon after this stage the pyroxene ceases to contain appreciable amounts of zirconium though the rock contains it abundantly. Zircon crystals are visible in thin sections of the transgressive hedenbergite granophyres and the acid granophyres where the average amounts of zirconium are 1400 and 850 ppm respectively. They have not been noted in the basic hedenbergite granophyres where the zirconium content is 250 ppm but no doubt minute crystals exist. The extreme insolubility of zirconium in acid magmas is clear from the separation of small early euhedral zircon crystals from granite, a point particularly commented upon by Vogt (1921, p. 634). Our results suggest that zircon separated from the Skaergaard magma when the zirconium

content was 300 or 400 ppm. The high values of 500 and 800 ppm given as the zirconium content of the two late liquids are because we have considered that the composition of the late rocks give the composition of the liquid. This is roughly true for most of the constituents, but for zirconium it cannot be true if, in fact, zircon is precipitated when the content of zirconium rises to 300 or 400 ppm. The reason suggested for this discrepancy is that the later liquids, produced by filter-press action, probably contained minute crystals of zircon which in thorough crystal fractionation should have accompanied the crystal precipitate but which, owing to their small size, remained in suspension in the liquid and give thereby a false figure for the zirconium content of the liquid. No doubt the solubility of zirconium in alkaline magmas is much greater than the figure we have suggested here for the solubility in granite magma.

14. Manganese

The results for manganese coming as they do partly from chemical analysis and partly from spectrographic determinations may not always be strictly comparable, and it is also not possible to discuss this element fully as we have not yet determined manganese in the iron ores.

Manganese is largely present in the pyroxenes and olivines, and no doubt also in the iron ores, but some is apparently present in the plagioclase structure. The proportion in all the minerals increases over the range of differentiation for which we have data except that there seems to be a slight fall in the latest plagioclase. The Mn in the olivine must be replacing in the Mg/Fe position but in the pyroxene it might be thought of as replacing either in the Mg/Fe or the Ca position or both. The fact that the amount in pyroxenes is about the same as in the olivines perhaps suggests that Mn replaces in both Mg/Fe and Ca positions equally easily. The Mn in the felspars is no doubt replacing Na and Ca. For the middle and late stages, the amount of Mn rises in successive differentiates at about the same rate as the Mn content of the liquid increases; and the ease of entry, as judged by the distribution factors, seems to change little. In the earliest plagioclase, of which two separate analyses were made, there is an indication of greater ease of entry than in the later. The smaller size of Mn relative to Ca, and the greater charge relative to Na, might be expected to produce concentration in the early stages. Consideration of the actual amounts in the plagioclases without reference to the amounts in the liquid would fail to give a satisfactory understanding of the Mn contents of the plagioclases.

In the original magma the amount of manganese present is 800 ppm. This is perhaps a little more than the average content of the hidden layered rocks (although the early border rocks in this case do not support this suggestion) and is slightly more than the amount in the hypersthene olivine gabbro 4077. The amount in later layered rocks is greater, reaching 3500 ppm in the fayalite ferrogabbros, after which it falls rapidly, becoming only about 150 ppm in the acid granophyres. The variation curve for manganese behaves in a closely similar way to that for Fe^{++}, as was shown in the earlier paper (1939, p. 227 and Plate 27).

15. Yttrium and Lanthanum

These two elements are concentrated in the later rocks. Yttrium rises above the sensitivity (30 ppm) only in the fayalite ferrogabbro and lanthanum only in the still later basic hedenbergite granophyre. Both elements enter the apatites abundantly and the late pyroxenes to a moderate extent. Yttrium is twice as abundant as

lanthanum on the average in the various minerals investigated. These elements are regarded as replacing calcium in apatite and pyroxene, and their absence from plagioclase is striking. In both the pyroxenes and apatites, the later fractions are richer than the earlier, although the reason for this is obscure. Hypotheses have been offered by Wickmann (1943, p. 488) and Nockolds and Mitchell (1948, p. 568) in explanation of rather similar data.

16. Strontium

Strontium is present in the plagioclases, and varies from 1000 to 5000 ppm; it is present in the apatites when they become primary precipitate minerals in roughly comparable amounts, but in the pyroxenes it is present only in insignificant amounts. Thus strontium can replace calcium in the plagioclase and apatite structures but not in the diopside structure. From an experimental study of certain artificial strontium silicates, Eskola (1922, p. 365) showed that a strontium anorthite exists but no strontium analogue of diopside, and he summarized an important conception as follows: "in those compounds in which the lime may possibly be replaced by magnesia and ferrous oxide it cannot be replaced by strontia or baryta." Over the range from An56 to An37 our figures indicate an enrichment in the Sr/Ca ratio. The actual amount of strontium also increases from the labradorite, An60 to the andesine An37 but falls in the last plagioclase analysed, An30. There is also a fall in the strontium content of the contemporaneous apatite. The late stage albite from a miarolitic cavity (Table 7) shows a still more marked reduction in the amount of strontium.

The melting point of strontium anorthite is known to be greater than 1700° C (Birch, Scharrer and Spicer, 1942) while calcium anorthite has a melting point of 1500° C, so that in the Sr–Ca anorthite series of solid solutions, the earlier should be strontium-rich and the later, calcium-rich as Eskola has shown (1922). On the basis of ionic size, however, it would be expected that the larger Sr ion would be less strongly bonded and that strontium anorthite would have the lower melting point, and thus be concentrated in the later solid solutions. In the plagioclase series it is of interest to note that the strontium seems to enter more readily into andesine than into either the earlier labradorite or the later oligoclase (see the distribution factors given in Table 6). The very small amounts of strontium in the albite from the miarolitic cavity probably indicate even less ease of entry in the extreme acid plagioclases, but this cannot be asserted definitely from our data as the content of Sr in the liquid from which this albite was formed is not known.

The amount of strontium in the original magma was 350 ppm (0·04% SrO); Noll (1934) gives 0·02% as the average for gabbros. Our somewhat higher figure might perhaps be expected as the original Skaergaard magma was richer in plagioclase than average gabbro. The amount of strontium precipitated in the early rocks seems to have been less than that in the liquid, and the content of successive residual magmas is estimated to have risen until about 60% of the magma had solidified. At this stage the hypersthene-olivine gabbros of the layered series were forming in which there are 600 ppm strontium. After this stage there is an irregular fall in the strontium content of the rocks and also in the estimated content of the residual magmas, and the final strontium content of the acid granophyre is only about 400 ppm. Noll's average data show a rise in strontium oxide content from 0·02% to 0·03% in passing from the gabbros to diorites and a fall to 0·01% in average granite,

so that the variation in the Skaergaard differentiates follows broadly that for the average calc-alkaline igneous rocks. Some of the variation in the strontium content of the Skaergaard rocks is due to the considerable variation in the amounts of plagioclase. Thus the reduction at the fayalite ferrogabbro stage is largely due to the low plagioclase content of this rock and not to any sudden change in strontium content of the plagioclase. On the other hand the strontium maximum in the late gabbros and ferrogabbros of the Skaergaard intrusion is due to the strontium content of the plagioclase felspars reaching a maximum at about An40, and it is likely that the high strontium content of average diorites, as found by NOLL, is due to the same cause.

17. Barium

This element, like strontium, is present in the felspars but it differs from strontium in not being present to any extent in contemporaneous apatites. The amount in the early plagioclases is about 50 ppm and it rises steadily to 600 ppm in the latest plagioclase (An30) of the layered series. The low content of only 30 ppm for the albite from a miarolitic cavity is of interest although this mineral is not on the direct line of Skaergaard crystal fractionation. While the barium content of the plagioclases rises strongly over the range shown by the layered series, the ease of entry as judged by the distribution factor does not change, the greater amount in the later plagioclases being due to the greater content in the liquid (*cf* Tables B and 6). The low amount of barium in the albite from the miarolitic cavity may indicate a reduction in the ease of entry for the latest stages, or may be due to all the barium in the liquid having been previously precipitated. ENGELHARDT (1936) found that barium is normally more abundant in the potash felspars than in the plagioclases, but that late potash felspar fractions, such as microcline from the Kragero pegmatites, have only 0·001 to 0·003% barium oxide. The factors involved in producing the low barium content of felspars of pegmatites are no doubt similar to those responsible for the low barium content of the albite from the Skaergaard miarolitic cavity.

The amount of barium in the original magma was 40 ppm, *i.e.* 0·005% BaO; ENGELHARDT gives 0·007% as the amount in average gabbro and anorthosite. With differentiation the amount in the Skaergaard rocks and the amount in the successive residual magmas rose. The general form of the curve for barium follows closely that for potassium but the amounts are approximately 100 times less (Figure 6). By the fayalite ferrogabbro stage the Ba content has reached 75 ppm and thereafter it rises to 400, 800 and 1400 in the successive granophyre fractions. In its rapid rise in these later filter-press fractions the behaviour of barium is in contrast with that of strontium, but is like that of rubidium, still to be considered. The large amount of barium in the later rocks is, no doubt, present in the potash felspars or perthites which we have not analysed. The large size of the barium ion is clearly the dominant factor in the accumulation of barium in the later rock fractions. The reduction of the barium content of very late fractions, which is indicated by the work of ENGELHARDT, and to some extent by that of NOCKOLDS and MITCHELL (1948, Figure 4), is not shown over the Skaergaard range of differentiates, although suggested by the analysis of the late stage albite of the miarolitic cavity.

18. Rubidium

This element is below the sensitivity (20 ppm) in all the minerals analysed except for the latest pyroxene, and here it may be that the rubidium is present in the small

amount of alteration products. In the rock series it only rises above the sensitivity in the granophyre differentiates where it is probably present in the acid felspar and also in micaceous late stage minerals.

XI. Comparisons with Other Rock Series and some General Considerations

1. The Skaergaard Trend of Differentiation

While 90% of the Skaergaard magma was solidifying, the trend of fractionation gave rocks of common types which are classed as gabbro-picrites, gabbros, *etc.* The solidification of the Skaergaard magma without fractionation also gave a common

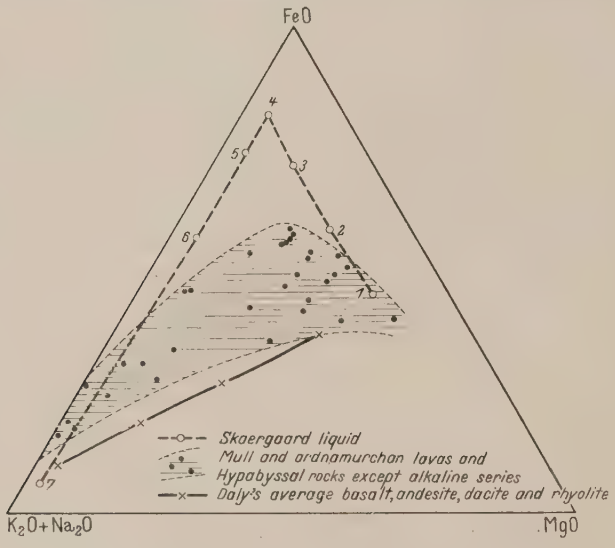

Fig. 10—The trend in composition of successive Skaergaard liquids from gabbro to ferrogabbro and granite compared with Daly's data for the calc-alkaline rock series and with analysed Mull and Ardnamurchan rocks (reproduced with permission from Meddel. om Grönland 1939 **165**).

type of olivine gabbro, the nature of which is indicated by the chilled marginal rock. Over much of the last 10% of the period of solidification, however, the uncommon rock type, ferrogabbro was produced while during the final stage a small amount, probably only 1%, of a granite fraction was formed. Although ferrogabbros as distinct rock masses are rare, Walker and Poldervaart (1949) have shown that a tendency towards development of ferrogabbro material is a normal result of the differentiation of basalt. By means of triangular diagrams, showing MgO, FeO and K_2O+Na_2O, of the type originally used to show the results of the Skaergaard differentiation (Figure 10), they have summarized the trend of differentiation for several basaltic provinces, including tholeiite as well as olivine basalt provinces (Kennedy's nomenclature 1933). They conclude (1949, p. 659) that "it may be regarded as established that iron enrichment is the normal trend of differentiation during the greater part of the crystallization of basaltic magma." Thus from the evidence of the Skaergaard intrusion and the more general considerations of Walker and Poldervaart, we regard it as established that strong crystal fractionation of basalt magmas of both olivine basalt and tholeiite types, may lead to rocks of

ferrogabbro composition. The variation in the amounts of the trace elements as found for the various Skaergaard fractions are therefore considered to have wide general significance.

When the major elements are considered, it is found that the ferrogabbro trend of fractionation of basic magma does not result in intermediate rocks of typical diorite composition and we shall show below that for several of the trace elements also, the composition of the intermediate rocks produced during the fractionation of the Skaergaard magma is unlike that of the diorites.

If the ferrogabbro trend is the result of the Skaergaard type of fractional crystallization of basalt, and the mugearite-trachyte trend, as exemplified by Hawaii (Macdonald 1949) is the result of another and perhaps more deep seated type, can the so-called normal calc-alkaline series of rocks be the result of yet another type of basalt fractionation? In the original account, possible reasons for the differentiation of basalt taking other trends than the ferrogabbro one and particularly the effect of a change in the state of oxidation of the iron, were considered (1939, pp. 310–313). It was concluded that strong fractionation of basalt magma under more oxidizing conditions would modify the proportions of the early phases separating but that eventually the liquid would reach a general ferrogabbro composition. We are now inclined to doubt this conclusion and to suggest that under more oxidizing conditions the residual liquid would develop a more ordinary andesitic composition, a hypothesis which is also considered possible by Walker and Poldervaart (1949, p. 661). However, even if andesitic rocks may be produced by fractionation of more oxidized basalt magma, the amount of such rocks would be relatively small, as is the case with ferrogabbros, and it still seems to us likely, as maintained in the original memoir (1939, p. 313), that the large masses of diorites and granodiorites of continental areas have resulted from other processes than crystal fractionation of basic magma at the time of emplacement.

In the Skaergaard intrusion strong fractionation of basic magma ultimately produced a relatively small amount of granitic liquid via a ferrogabbro stage and it is interesting to consider whether this provides any clue to the much discussed problem of the origin of granites. From the small production of granitic magma by fractionation in the Skaergaard intrusion and from much other petrogenetic work, (*cf* Bowen, 1947, p. 273) the possibility of the production of some granite by fractional crystallization of basalt is clearly shown. When however the vast granitic and granodioritic masses of continental regions are considered, it seems unlikely that they have originated at the time of emplacement or mobilization, by fractionation of basic rock. The hypothesis favoured in the original paper (1939, pp. 324–335) of the origin of the granites and granodiorites by re-melting of the sial layer, combined with some differentiation, still seems the most satisfactory. The general arguments for this view have been effectively stated by Kennedy (1938), Daly (1949) and others.

If the origin of the granitic masses in the upper crust be by re-melting of pre-existing material of the sial layer combined with only slight fractionation, the problem of the actual origin of granite is pushed back in time to the origin of the sialic part of the lithosphere. It seems likely that the sial layer was produced by some sort of fractional crystallization process from an originally homogeneous silicate liquid having a general basic composition. The small amount of granitic liquid in the Skaergaard intrusion, developed by crystal fractionation from a basic rock, perhaps provides a legitimate analogy with this original differentiation process but

whether this be so or not, we shall show below that there is a similarity between the general trace element composition of the granitic rocks of the lithosphere and the Skaergaard acid differentiate which definitely resulted from fractional crystallization.

2. Comparison of the Trace Elements in the Skaergaard and Other Rock Series

Comparisons of the trace element composition of the successive Skaergaard differentiates may be made with the Caledonian igneous suite of Scotland (Nockolds and Mitchell, 1948) and with the igneous rocks in central Roslagen, Sweden (Lundegårdh, 1946). Further highly significant comparisons can also be made with the average data for ultrabasic, basic, intermediate and acid rocks as obtained by Goldschmidt and other workers.

For the Caledonian igneous rocks the most clearly defined trend is that from pyroxene mica-diorite to granite and this is considered by Nockolds and Mitchell (1948, Figure 3) to be the line of liquid descent, due to crystal fractionation. This series may be compared with the later Skaergaard differentiates from the basic hedenbergite granophyre onwards, that is, with the right hand part of our Figure 3 from about 95% to 100% solidified (for the major constituents over this range see 1939, pp. 212–213 and Figure 40). Where most closely comparable with the Caledonian series, data for the Skaergaard differentiates are unfortunately meagre.

When comparison is attempted between the earlier rocks of the Skaergaard intrusion and the Caledonian series, the extended curves available for the Skaergaard intrusion from 0–95% solidified should be compared with the rather bunched points for the accumulative rocks of Nockolds and Mitchell. Over this range the best way of making the comparisons is to use averages for broad groups, classified under the headings: ultrabasic, basic, intermediate and acid (Table 15).

Average data for the trace element composition of ultrabasic, basic, intermediate and acid rocks of varied provenance are also available from the work of Goldschmidt and his collaborators, Strock, Noll, and Engelhardt, and from Sandell and Goldich, and Tröger; these data are summarized in Table 15. In this table we have given the average Caledonian data separately as they were obtained from a series of related rocks and by the same technique as for the Skaergaard example.

The Skaergaard figures in the ultrabasic column (Table 15) are for the gabbro-picrite which is not a true ultrabasic rock but intermediate between peridotite and gabbro; this results in some understandable discrepancies from the Caledonian and average data. In the column for basic rocks, two values for the Skaergaard intrusion are given, the first is for the hypersthene-olivine gabbro of the layered series which resulted from a process of crystal sorting; the second, in brackets, is for the original Skaergaard magma which represents the composition of a completely liquid magma of general gabbroic composition. In obtaining the general average for basic rocks no distinction between rocks due to crystal sorting and those on the liquid line of descent was made and thus for comparison the average of the two Skaergaard figures may fairly be used. For the Skaergaard representatives of the intermediate rocks, the average values for the ferrogabbros, in which the felspars are andesine and thus comparable with the diorites are given; figures are also given in brackets for the basic hedenbergite granophyre in which the silica percentage is comparable with that of the intermediate rocks. To represent the Skaergaard acid rocks, average figures for the differentiates with silica percentages ranging from 59 to 75 are given first, and these are followed by figures in brackets for the acid granophyres by themselves; the latter are the more typical acid rocks (SiO_2 75%) and thus give a

Table 15—Trace elements in Skaergaard differentiates compared with those in certain Caledonian plutonic rocks of Scotland and with average values taken from papers by Goldschmidt and others

Figures in parts per million.

		Ultrabasic	*Basic*	*Intermediate*	*Acid*	*Average igneous material*†
P	Skaergaard	90	260(280)	5000(3000)	1500(80)	
	Caledonian	—	1000	1200	750	800
	Average (TRÖGER)	**400**	**1500**	**1300**	**800**	
Ga	Skaergaard	8	19(17)	20(28)	33(30)	
	Caledonian	3	17	20	40	15
	Average (G)	—	**8**	—	**15**	
Cr	Skaergaard	1500	230(120)	<1(<1)	11(14)	
	Caledonian	2500	300	150	60	200
	Average (G)	**5000**	**500**	**100**	**3**	
V	Skaergaard	120	220(140)	4(<1)	9(18)	
	Caledonian	100	150	100	80	150
	Average (TRÖGER)	**130**	**250**	**130**	—	
Ti	Skaergaard	900	5000(8500)	15000(7000)	700(1800)	
	Caledonian	4500	5000	7000	3000	4400
	Average (TRÖGER)	**900–8000**	**10000**	**8000**	**3500**	
Li	Skaergaard	2	2(2)	3(15)	20(13)	
	Caledonian	10	30	35	20	65
	Average (STROCK)	**2**	**10**	**20**	**150**	
Ni	Skaergaard	1000	120(170)	<2(4)	9(8)	
	Caledonian	700	200	80	20	100
	Average (G)	**4000**	**200**	**50**	**3**	
Co	Skaergaard	90	48(53)	20(12)	5(5)	
	Caledonian	200	70	30	15	40
	Average (G)	**300**	**100**	**40**	**10**	
Cu	Skaergaard	100	67(130)	400(500)	200(20)	
		—	—	—	—	100
	Average (S & G)	—	**150**	**40**	**16**	
Sc	Skaergaard	<10	20(12)	10(8)	<10(<10)	
	Caledonian	20	20	10	<10	5
	Average (G)	—	**20**	**5**	**1½**	
Zr	Skaergaard	30	33(50)	20(250)	1200(850)	
	Caledonian	<10	60	150	200	220
	Average (TRÖGER)	—	—	—	**300**	
Mn	Skaergaard	1200	700(800)	3300(2000)	800(150)	
	Caledonian	400	1100	750	350	1000
	Average (TRÖGER)	**200**	**1500**	**1000**	**1000**	
Sr	Skaergaard	100	600(350)	400(400)	450(350)	
	Caledonian	20	1000*	1300*	1300*	150
	Average (NOLL)	—	**180**	**250**	**90**	
Ba	Skaergaard	10	18(43)	60(400)	1100(1400)	
	Caledonian	40	150	700	1300	250
	Average(ENGELHARDT)	**3**	**70**	**300**	**450**	
S	Skaergaard	—	450(500)	2500(450)	—	
	Caledonian	—	—	—	—	520
	Average (TRÖGER)	**2500**	**2000**	**1000**	**400**	

1. For the Skaergaard intrusion the figures are averages from Table F:—
Figures in the *ultrabasic* column are for gabbro-picrite (Column A). (This is not a true ultrabasic rock but intermediate between ultrabasic and basic.)
Figures in the *basic* column are for hypersthene-olivine gabbro (Column B). Figures in brackets are the values for the average chilled marginal gabbro (the original Skaergaard magma).
Figures in the *intermediate* column are for ferrogabbros, the average of Columns D, E and F. Figures in brackets are for the basic hedenbergite granophyres SiO_2 50% (Column G).
Figures in the *acid* column are for transgressive granophyres, SiO_2 59% to 75% (mean of columns H & I). Figures in brackets are for the acid granophyres only (Column I).
2. For the Caledonian plutonic rocks the figures are for the Garabal Hill-Glen Fyne Complex (NOCKOLDS and MITCHELL 1948 Table II):—
Figures in the *ultrabasic* column are average of columns 1–3.
Figures in the *basic* column are average of columns 4–8.
Figures in the *intermediate* column are average of columns 9–14.
Figures in the *acid* column are average of columns 15–21.
3. The average data are taken from GOLDSCHMIDT (1934, 1937 and 1938), NOLL (1934), TRÖGER (1935), STROCK (1936), ENGELHARDT (1936) and SANDELL and GOLDICH (1943).

* Figures for Sr in the Caledonian rocks indicated with an asterisk are probably too high.
† After GOLDSCHMIDT.

fairer comparison with the general data for the acid rocks. In making comparisons, too much stress should not be placed on the absolute values, but more on the general trends.

For the ultrabasic and basic rocks, the amounts of the trace elements in the Skaergaard, Caledonian and average rocks are reasonably comparable in most cases, especially when it is remembered that the figures for the Skaergaard intrusion in the ultrabasic column are really for rocks transitional to the basic rocks. Phosphorus is one of the elements showing a considerable discrepancy, due probably to the fact that the Caledonian and average rocks include types on the liquid line of descent. The data for manganese also show considerable variation, the reason for which is not clear.

For the intermediate rocks the amounts of many of the trace elements are moderately similar if the average of the two figures given for the Skaergaard intrusion be taken, but for chromium, vanadium, nickel, copper and to some extent cobalt the values are widely different. Chromium, vanadium and nickel are reduced to very small amounts in the intermediate rocks of the Skaergaard intrusion while they remain in much higher amounts in the Caledonian and average intermediate rocks which are on the average diorites. Cobalt shows the same tendency as nickel, but less strongly. In the case of copper, the marked enrichment in the Skaergaard intermediate rocks is in contrast with the low amounts in average intermediate rock. Although in the Skaergaard case an immiscible sulphide liquid probably formed, the copper enrichment is independent of this.

For the acid rocks, there is moderate agreement between the various sets of figures, especially if the average of the two figures given for the Skaergaard is taken. The causes of such discrepancies in the figures as can be regarded as significant will not be considered here; for some cases, general explanations will occur readily enough to the reader while to attempt detailed consideration would involve too many doubtful assumptions.

Judged by the Skaergaard intrusion, a process of fractional crystallization of basic magma results in the chromium, vanadium and nickel values in the intermediate rocks being reduced to very low amounts. Even if it is possible for some modification in the trend of fractional crystallization of basaltic magma to yield the much more abundant andesitic or dioritic type of intermediate rock in the middle stages, it is still likely that the process would result in these three elements being reduced to low values because there would still be early ferromagnesian minerals into which chromium, vanadium and nickel would enter: with effective fractionation the result should be the almost complete elimination of these elements. If fractionation were not thoroughly effective then it would not result in the production of intermediate and acid rocks. Furthermore if hornblende or biotite were the ferromagnesian minerals concerned instead of olivine and pyroxene, it seems to us that fractionation should still result in the elimination of the chromium, vanadium and nickel in early hornblende and biotite. Thus without strong fractionation, dioritic and granitic rocks could not be produced from basic magma and with strong fractionation, the amounts of chromium, vanadium and nickel should be reduced to low values (*cf* Lundegårdh, 1949, p. 6 and 11). Since in average diorites these elements are not reduced to really low values the trace element evidence favours the view that the bulk of such intermediate rocks have not been produced by fractional crystallization of basaltic magma. Although the bulk of intermediate rocks have a higher chromium, vanadium and nickel content than would be expected if produced by fractionation from basalt magma it may

be that future investigation will reveal small amounts of special types of diorite and andesite having low amounts of these three elements, and therefore possibly produced by direct fractionation of a rather oxidized basalt magma.

With further work it may be found that the intermediate rocks will fall into two types having contrasted trace element composition :

1. Rocks containing very low amounts of chromium, vanadium, and nickel, and produced by fractionation of basic magma, *e.g.* the ferrogabbros, the basic hedenbergite granophyres, and, perhaps also the mugearites and the special types of diorite and andesite just mentioned.

2. The typical diorites with markedly higher chromium, vanadium, and nickel content, and probably having some other origin than by fractionation of basic magma. The latter may be the more basic differentiate produced by fractionation of granodiorite magma while granite is the more acid fraction (*cf* NOCKOLDS, 1941). Such a granodiorite magma containing chromium, vanadium and nickel in small amounts would be likely to give an early fraction having higher values of these elements than intermediate rocks produced by strong fractionation of basic magma. On the other hand, the residual granite from fractionation of granodiorite magma would be depleted in these constituents like the residual granite from fractionation of basic magma.

While the common intermediate rocks may be derived from a granodiorite parent magma, and another, less abundant group may be derived from the fractionation of basic magma, there is apparently a third group, the importance of which is far from clearly evaluated, produced by mixing of acid and basic material. The values for Cr, V and Ni determined for the dioritic intermediate rocks are the same as would be obtained if basic and acid rocks were mixed in about equal proportions and this is the proportion required to give the appropriate composition of the intermediate rocks in terms of the major constituents. Such a mixing process* would also give satisfactory values for the other trace elements investigated. Thus the trace element data, used in this broad way would not distinguish between intermediate rocks resulting from a mixing process and those produced as an early fractionate from sialic material.

GOLDSCHMIDT considered that the trace element composition of granites is what would be expected on crystal chemical grounds if they were a late liquid resulting from fractional crystallization. This view is supported by the evidence from the Skaergaard intrusion as the granite residuum, produced by the strong fractionation of the basic magma, is reasonably similar in trace element composition to that of granites. The Skaergaard granite residuum resulted from the contemporaneous fractional crystallization but we would think of the bulk of normal granites as essentially the result of fractional crystallization in the early stages of the earth when the granite shell was formed. The origin of granites and diorites cannot be solved by such flank attacks as can be made in this place but we believe that consideration of the trace element composition of the rocks will play an important part in the main attacks on these problems.

The way in which certain trace elements have behaved under the conditions of strong fractionation shown by the Skaergaard intrusion is of interest in relation to the occurrence of these elements in mineral deposits. GOLDSCHMIDT has pointed out how a trace element having similar ionic size and charge to a major constituent is *camouflaged* in the minerals of that major element. An extreme case is hafnium in zirconium minerals, and another frequently quoted case is gallium in aluminium

* As previously mentioned (1939, p. 321) convection currents may well provide the mechanical means by which mixing processes might become effective.

minerals. In a rather similar way it is obvious that, when crystal fractionation of basic magma produces little effective concentration of an element because its properties are such that it readily enters one or other of the crystal phases formed, then, even though it be a fairly abundant element, it has a much reduced chance of giving rise to mineral deposits of magmatic origin. Because scandium enters readily into the pyroxene structure, it does not accumulate in residual magmas during fractionation and does not give rise to scandium mineral deposits. Vanadium and cobalt, which also have no marked tendency to become stored up in the residual magmas nor any strong tendency to be precipitated in early mineral phases, are rarer in mineral deposits of magmatic origin than chromium and nickel on the one hand, which tend to enter early mineral phases, and zinc, manganese and copper on the other hand, which do not enter so readily into the minerals of the early or middle stages of the fractionation process but tend to be left in the late fractions. If there is a development of an immiscible liquid sulphide phase these generalizations will not hold without qualifications for chalcophile elements. Among the trace elements of large ionic size, strontium tends to separate abundantly in the intermediate felspars, and therefore, does not become strongly concentrated in the residual liquids, while barium tends to remain in the residual magma, at any rate until a late stage. This may be an important factor in making barium* a more abundant element than strontium in many mineral deposits of presumed ultimate magmatic origin despite it being less abundant in terrestrial matter as a whole (see last column of Table 15).

In the Skaergaard intrusion chromium, vanadium and nickel, after being present in very small amounts in the intermediate stages, increase somewhat in the latest residual liquids, and the reason for this is not understood; the amounts of these elements in the granitic stage of the Skaergaard intrusion are surprisingly close to the amounts in average granite.

There is a final point connected with the problem of the origin of the granites which it seems proper to make. The successful crystal chemical explanation of the average trace element composition of igneous rocks which Goldschmidt presented is based on the assumption that crystal fractionation is the primary cause of the main rock series. In particular, he has suggested that granite is a late residuum from such a fractionation process. The Skaergaard data, by providing a particularly clear, special case, support Goldschmidt's views. The alternative hypothesis of the origin of granites by metasomatic processes acting on a variety of rock types would, in our opinion, produce a more varied trace element composition, and one which would not be expected to have the trace element features of a residual liquid formed by fractional crystallization processes.

In supporting the hypothesis that the granites are on the whole the result of crystal fractionation, we do not suggest that the large granite intrusions of the upper crust have been derived, at the time of their emplacement, by the differentiation of basic magma, but only that the granite material was so derived originally, and that at the time of its final emplacement it was merely rendered mobile again by a process of re-fusion, or partial re-fusion. This view implies that re-fusion did not alter intrinsically the trace element composition of the granite material, which remained essentially the same as when originally produced as a sial crust by fractional crystallization during the early stages of cooling of the planet.

* Barytes deposits having cone-sheet form and apparently a late differentiate of the Alno alkaline rocks have recently been described (Eckermann, 1948).

Appendix

Inclusions and alteration products in the analysed minerals and their effects on the analytical data

The existence of impurities of other minerals in small amounts can be seen on close scrutiny of the mineral grains separated for analysis. The impurities are visible more clearly in thin sections of the rock or when the samples for analysis are finely crushed and examined under high magnifications.

The inclusions are of various types which may be classified as follows:

I. Small crystals of one or other of the primary minerals, incorporated while the larger crystals were forming. These are rather abundant in the outer parts of some of the crystals.

II. Minute crystals of apparently high temperature minerals distributed thinly over curved planes which appear to be sealed planes of fracture.

III. Material filling irregular fractures or cleavage planes and apparently consisting of low temperature, hydrated minerals such as honblende, serpentine, chlorite and sericite.

IV. Acicular, platey or rounded inclusions, usually with special orientation and arranged along certain planes in the crystal. These are considered to be due to ex-solution.

Inclusions in crystals were described in a classic paper by Sorby (1858); later work was done by Judd, Williams and others and Teall summarized the position up to 1888 in his " British Petrography " (1888, pp. 16–31). Judd figured inclusions classed here as type II (1885, Plate X, Figures 3, 4, 5) and also inclusions of type IV (1885, Plate X, Figure 6 and later figures) but the descriptions seem prejudiced by his theory of their origin by the action of solvents under pressure. No comprehensive account since then has come to the notice of the present authors, although A. G. MacGregor (1931) has described the clouding of felspars by swarms of inclusions produced by thermal metamorphism, and Ingerson (1947) has recently re-used Sorby's method for determining the temperature of formation of minerals by examination of liquid inclusions*. The general problem of inclusions in the crystals of igneous rocks deserves renewed attention; here, however, we wish only to consider what effect they may have on the trace element mineral compositions given in this paper.

Inclusions of type I are generally of fair size and would only rarely be present in our hand-picked material. They consist of the usual minerals of the rock and would have the same trace element composition. From the point of view of the mineral analyses they would be the same as accidental particles of the other minerals attached to the hand-picked specimen. A maximum possible value for contamination of this sort can usually be fixed by comparison of the trace element data for the rock and its separate minerals (p. 137).

Inclusions along healed fractures (type II) consist of minute crystals, which are sometimes well shaped and sometimes ragged, and usually have a greatest dimension less than 0·005 mm so that many may occur in the thickness of a slice. These inclusions are thinly scattered over slightly curving surfaces cutting through

* Since this was written an important paper by Tuttle (1949) on planes of liquid inclusion in quartz has appeared. These resemble in many ways the healed fracture type of inclusion (II above) and a rather similar explanation of their origin is given (p. 334).

the crystals. The form of the surfaces defined by these inclusions indicates that they were once fracture surfaces, although the fractures do not now exist. In the felspars of the Skaergaard intrusion the sealed fractures seldom follow the cleavage directions. It is tentatively suggested that at an early stage, when perhaps 5 or 10% of magma still occupied the space between the precipitate of crystals, fractures were produced (perhaps by pressure of one crystal on another due to the thickness of the pile of loose crystals) and that these were filled by the still liquid magma occupying the crystal interstices. On crystallization of the injected liquid some would form the same material as the host, and by growing in optical continuity would seal the original crack, while some would form solid material different from the host and give rise to the thinly scattered crystals which define the position of the initial fracture. In the plagioclases the crystals along the sealed cracks have high refractive indices and seem to be pyroxene and olivine; in the olivines the minerals scattered along the sealed cracks can often be seen to have a lower refractive index than the host and are thought to be composed of felspar and pyroxene. In rare cases small cavities occur at intervals along the line of the sealed cracks which also help to define their position.

The type III included material is that which is commonly observed along fractures when thin sections of plutonic rocks are examined. This type cuts the sealed fracture type and was thus produced later. In the felspars the type III inclusions follow cleavage directions more often than type II suggesting that with lowering temperatures the felspars fractured more readily along the cleavage. In type III the cracks are occupied by material different from the host, and thus are not sealed in the same way as type II. The minerals occupying the fractures are lower temperature, hydrated minerals produced, in part, by reactions between the solutions and the host crystal.

The latest cracks visible in thin sections are irregular or cleavage cracks produced during preparation of the slices and filled with Canada balsam.

An orientated intergrowth of ortho and clino-pyroxene is common in many of the Skaergaard rocks and is regarded as due to ex-solution (1939, pp. 81–83). In the olivines and pyroxenes and sometimes in the felspars there are small orientated needles or plates of foreign material (type IV inclusions) which are presumably also the result of ex-solution. Such orientated inclusions are a common feature of the minerals of plutonic rocks and they were frequently commented upon in the early days of microscopic petrology, particularly by Judd (1885). In some of the felspars of the Skaergaard layered series, minute orientated rods of opaque material occur while in others they are absent, and no reason for the erratic distribution has yet been found*. The orientated inclusions of the plagioclases probably contain some of the ferric iron shown to be present by analysis. It is considered that the material of the orientated inclusions formed part of the crystal structure at the temperature of formation and that it separated out on cooling because it was not so readily accommodated in the crystal lattice at the lower temperatures. The cooling of large plutonic masses is necessarily slow and there is much time for such ex-solution effects to take place.

* It would be interesting to consider the general distribution among igneous rocks of these types of inclusions, but this cannot be attempted here. The sealed fracture type are better displayed in the Skaergaard rocks than in the few other examples of gabbros that we have examined. In certain of the Skye gabbros the material along the healed cracks seems to be hornblende and thus forms a transition from type II to type III inclusions. Type II inclusions are clearly rare among the phenocrysts of lavas if they occur at all. In considering inclusions Judd noticed differences between plutonic rocks and high level dykes and sills and lava flows.

The inclusions which are due to ex-solution should consist of elements having some special relation to the host and thus should not be regarded as impurities from point of view of the high temperature crystal-chemistry of the minerals. On the other hand the material along the sealed fractures is extraneous matter which in our view had the general composition of the residual liquid present interstitially at the time of formation of the fractures.

In so far as the material forming the sealed fracture type of inclusion is reasonably near the composition of the analysed minerals of the rock, an upper limit to the amount present in any sample may be set, as for the accidental fragments, by the method of internal comparison (see p. 137); in so far as the material has a special composition, because the residual liquid from which it came is a more extreme product of fractionation, this method of estimating the effect on the analysis cannot be used.

The material present as type III inclusions is more variable in nature and amount than that of type II and its effect in our analyses is unfortunately more unpredictable. The liquid injected into cracks which deposited, or produced by reaction with the host, the type III inclusions would seem to have been a late stage watery solution, no doubt containing trace elements in very different quantities from the original magma. If a trace element appears in small, and about equal amounts in all the chief minerals of a rock, and if it is also the sort of element which might be expected to become concentrated in residual solutions, then it may well be that the element is not in the crystal structure but is present in type III inclusions or alteration products. Of the elements considered in this paper, lithium shows the above characteristics in the minerals of the early and middle stages, and we suggest that it may be present as inclusions or alteration products due to the injection of a late stage residual liquid along cracks in the crystals. In certain late pyroxenes and olivines there is considerably more lithium present and then no doubt it is also entering the crystal structure, replacing ferrous iron.

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BOOK REVIEW

Correns, W.; **Einführung in die Mineralogie.** Pp. viii + 414. 405 illustrations and 1 plate. Springer Verlag, Heidelberg 1950. Price DM.41·60.

Mineralogy is concerned not merely, or even mainly, with the characters of minerals, a study which might properly fall within the domains of physics and chemistry, but also with the mode of formation and natural association of minerals as elements in the structure of the earth's crust. This fact is clearly recognized by Professor Correns and forms the basis on which the present book is designed. The latter is a volume similar in certain respects to Eskola's "Krystalle und Gesteine" and in its contents is unlike anything at present available in the English language. The book is based on the general lectures developed by Professor Correns over a period of some twenty-two years and, as one might expect from a teacher of his reputation, bears all the marks of deep thought and careful selection.

The first part of the book follows more or less conventional lines and deals lucidly with morphological and mathematical crystallography, with crystal chemistry and with crystal physics. Within this section it is notable that various aspects of the subject are more fully and carefully explained than is normally the case in textbooks of mineralogy. At the same time there is no tendency to overload the text with unnecessary detail.

Part II deals with the natural occurrence and association of minerals, and presents a brief but comprehensive, and exceedingly valuable dissertation on the theoretical basis of petrology. In this petrological section there is an initial discussion on the physical chemistry of rocks followed by a more detailed consideration of magmatic, sedimentary and metamorphic processes of rock formation, and geochemistry in its wider sense.

Part III consists of a series of tables to which are relegated certain crystallographic data together with valuable information concerning the characters of the more important minerals and rocks. Apart from the tabular information there is no systematic description of mineral species along conventional lines.

The entire work is admirably conceived, and constitutes a real and valuable introduction to the entire science of minerals. It is only to be regretted that such a book is not more widely accessible to English speaking students.

W. Q. Kennedy (Leeds)

Table A—Major and trace constituents in various specimens of the fine grained marginal gabbro, and comparisons

	r	s	*I* 1825	*II* 1724	*III* 1922	*IV* 4093	*Y*	*M*	*N*	*O*	*P*	*Q*	*V* 1093	*VI* 1112	*VII* 1057
			Major constituents as weight per cent of the oxide												
SiO_2	0·39A		48·01	47·83	—	—	**47·92**	47·54	—	48·24			47·26	47·81	48·89
Al_2O_3	0·57		19·11	18·62	—	—	**18·87**	13·91	—	17·88			14·17	13·64	13·60
TiO_2	0·64		1·51	1·29	—	—	**1·40**	2·60	—	·97		1·70	2·69	2·51	2·64
Fe_2O_3	0·67		1·20	1·16	—	—	**1·18**	4·59	—	3·16			4·68	4·50	4·72
MgO	0·78		7·72	7·92	—	—	**7·82**	6·58	—	7·51			6·65	6·51	4·54
FeO	0·83		8·44	8·87	—	—	**8·65**	8·58	—	5·95			8·32	8·84	9·65
Na_2O	0·98		2·34	2·54	—	—	**2·44**	2·42	—	2·55			2·30	2·54	2·74
CaO	1·06		10·33	10·59	—	—	**10·46**	10·97	—	10·99			10·64	11·29	10·09
K_2O	1·33		·17	·20	—	—	**·19**	·58	—	·89			·81	·35	·30
			Trace constituents as parts per million of the element												
P^{+5}	0·34		(300)	(260)	—	—	**(280)**	(1000)	—	—	—	1500	(1000)	(1050)	(1700)
Ga^{+3}	0·62	1	20	15	15	15	**17**	20	20	8	—		20	20	20
Cr^{+3}	0·64	1	100	200	200	700	**170**	175	350	350	—	700	200	150	30
V^{+3}	0·65	5	60	150	200	200	**140**	200	250	—	—	250	200	200	200
Mo^{+4}	0·68	1	*	*	*	*	*	*	*	—	1		*	*	*
Li^{+1}	0·78	1	3	2	2	8	**2**	4	6	10	—		5	3	5
Ni^{+2}	0·78	2	200	150	150	250	**170**	75	70	160	100	400	80	70	30
Co^{+2}	0·82	2	60	50	50	50	**53**	40	25	80	40		40	40	25
Cu^{+2}	C.0·8	10	80	150	150	150	**130**	150	120	—	120	250	150	150	300
Sc^{+3}	0·83	10	10	*	20	30	**12**	10	15	20	—		10	10	*
Zr^{+4}	0·89	10	50	30	70	30	**50**	70	80	—	—		70	70	100
Mn^{+2}	0·91	10	(920)	(690)	—	—	**(800)**	(1600)	—	—	—	1500	(1450)	(1700)	(1850)
Y^{+3}	1·06	30	*	*	*	*	*	*	*	2	—		*	*	*
La^{+3}	1·22	30	*	*	*	*	*	*	*	8	—		*	*	*
Sr^{+2}	1·27	10	250	300	500	200	**350**	200	350	170	—	170	200	200	200
Ba^{+2}	1·43	5	10	40	80	20	**43**	20	200	60	—	250	20	20	40
Rb^{+1}	1·49	20	*	*	*	*	*	*	*	18	—		*	*	*
S			(400)	(630)	—	—	**(500)**	100	—	—	—	2000	(100)	(100)	(200)

The new data for trace elements determined spectrographically except where figure is in brackets when the values given are from chemical analysis.

An asterisk indicates that the element is present in amounts less than the sensitivity. A dash indicates absence of data.

The following elements were also sought but are not present in amounts above the sensitivity shown in brackets: B(1), Be(10), Ge(10), Sn(5), Zn(300), Sb(300), Ce(500), Cd(300), Ag(1), Pb(10), Tl(50).

r Ionic radius in Ångstrom units at valency shown.

s Sensitivity by the spectrographic method employed for the new data.

I 1825 Fine grained olivine gabbro, eastern margin (see WAGER and DEER, 1939, p. 137).

II 1724 Fine grained olivine gabbro, southern margin (1939, p. 138).

III 1922 Fine grained olivine gabbro, western margin (1939, p. 138).

IV 4093 Fine grained olivine gabbro, northern margin (1939, p. 139).

Y Average of I, II, and III taken to represent the composition of the original Skaergaard magma.

M Average for two of the olivine basalts (columns V and VI).

N Average for two Karroo dolerites (See WALKER and POLDERVAART, 1949, p. 644, columns 17 and 5).

O The major constituents are for average gabbro as computed by DALY (1933 p. 17). The trace constituents are values obtained by GOLDSCHMIDT and others for a composite gabbro sample made by NOLL from eleven typical gabbros. Additional data for this sample are:—B 3, Be 1, Ge 1, Sn 10 ppm.

P Amounts in average basic rock with silica percentage 48 taken from curves given by SANDELL and GOLDICH (1943) and based on chemical analyses. Additional data from this source are Pb 4, Zn 100 ppm.

Q Average values for gabbro after TRÖGER (1935, p. 316).

V 1093 Olivine basalt, Blosseville Coast, East Greenland (see WAGER, 1934, pp. 32–34). The spectrographic data are new.

VI 1112 Olivine basalt, Blosseville Coast, East Greenland (WAGER, 1934, pp. 32–34). The spectrographic data are new.

VII 1057 Tholeiitic basalt, Blosseville Coast, East Greenland (WAGER, 1934, pp. 32–34 and WAGER, 1935, pp. 14–17). The spectrographic data are new.

Table B—Major and trace constituents in minerals from the early rocks of the border group and from the layered series

	r	s	*Plagioclase* A I 1851	B II 4077	C III 3662	D IV 2580 & 1907	E V 4145	F VI 4142	*Pyroxene*[1] A VII 4289 & 1846	B VIII 4077	C IX 3662	D X 1907	E XI 4145	F XII 1881	*Olivine* A XIII 1846	B XIV 4077	C	D XV 1907	E XVI 4146 & 4145	F XVII 4139 & 1881	*Ilmenite* B XVIII 2308	C XIX 3662	D XX 1907	E XXI 4145	*Magnetite*[2] B XXII 2308	C XXIII 3662	D XXIV 1907	E XXV 4145	F XXVI 4142	*Apatite* D XXVII 3649	F XXVIII 4142	
Major constituents as weight per cent of the oxide																															
SiO_2	·39		—	54·35	—	58·03	59·03	—	50·39	49·89	47·92	49·38	46·06	42·62	—	38·11		33·72	31·85	30·15											SiO_2
Al_2O_3	·57		—	28·35	—	25·81	25·02	—	3·54	3·74	4·60	2·93	4·06	5·24	—	nil		nil	tr	·07											Al_2O_3
TiO_2	·64		—	—	—	—	—	—	·87	·93	1·71	0·71	1·58	1·69	—	tr		tr	·01	·20											TiO_2
Fe_2O_3	·67		—	·49	—	·38	·75	—	1·95	·21	·71	1·48	1·21	3·74	—	·15		·05	·11	·43											Fe_2O_3
MgO	·78		—	·02	—	·04	·02	—	15·82	14·59	12·03	10·66	4·51	·47	—	30·50		18·07	8·49	1·05											MgO
FeO	·83		—	·18	—	·30	·45	—	8·47	12·81	18·85	17·36	24·02	31·54	—	31·48		47·91	58·64	65·02											FeO
Na_2O	·98		—	4·95	—	6·47	6·73	—	·70	·43	·51	0·37	·53	1·02	—	—		—	—	—											Na_2O
CaO	1·06		—	11·19	—	8·01	7·14	—	18·41	16·83	12·95	16·83	17·37	12·27	—	·02		nil	·18	2·18											CaO
K_2O	1·33		—	·38	—	·46	·46	—	·14	·13	·13	0·11	·21	·23	—	—		—	—	—											K_2O
Trace constituents as parts per million of the element																															
Ga^{+3}	·62	1	35	50	50	50	40	200	3	3	*	5	10	10	*	3		5	5	5	5	3	3	5	20	30	30	60	30	*	3	Ga^{+3}
Cr^{+3}	·64	1	10	*	*	*	*	*	3000	350	*	*	*	*	20	*		*	*	*	2	*	*	*	200	*	*	*	*	*	3	Cr^{+3}
V^{+3}	·65	5	10	*	10	10	10	*	300	250	100	30	*	*	*	*		*	*	*	600	300	20	*	2000	800	300	*	*	*	*	V^{+3}
Mo^{+4}	·68	1	*	*	*	*	*	*	*	3	5	*	*	*	3	10		10	10	30	3	*	3	3	5	3	3	3	3	*	*	Mo^{+4}
Li^{+1}	·78	1	2	3	3	3	3	*	3	2	*	*	2	50	2	3		5	5	15	*	*	*	*	*	*	*	*	*	*	*	Li^{+1}
Ni^{+2}	·78	2	10	*	*	*	*	*	200	140	50	*	*	*	2000	325		10	*	*	150	50	*	*	300	100	*	*	*	*	*	Ni^{+2}
Co^{+2}	·82	2	*	*	*	*	*	*	60	50	60	40	30	15	150	125		100	50	20	80	100	70	30	80	80	60	30	3	*	*	Co^{+2}
Cu^{+2}		10	15	35	15	250	700	10	*	35	100	300	1000	100	20	20		300	400	20	100	30	300	400	20	50	500	300	400	200	*	Cu^{+2}
Sc^{+3}	·83	10	*	*	*	*	*	*	80	30	50	150	80	80	*	*		*	*	*	*	*	*	*	*	*	*	*	*	10	30	Sc^{+3}
Zr^{+4}	·89	10	*	*	*	*	*	30	30	50	30	50	*	*	*	*		*	*	*	*	*	*	*	*	*	*	*	*	20	30	Zr^{+4}
Mn^{+2}	·91	10	100	C.20	50	50	200	100	1500	(2100)	(2700)	(1700)	(3250)	(6000)	2000	(1700)		(3200)	(6600)	(7800)	5000	—	—	—	5000	—	—	—	—	400	400	Mn^{+2}
Y^{+3}	1·06	30	*	*	*	*	*	*	*	*	*	*	*	500	*	*		*	*	*	*	*	*	*	*	*	*	*	*	1000	4000	Y^{+3}
La^{+3}	1·22	30	*	*	*	*	*	*	200	*	*	*	150	300	*	*		*	*	*	*	*	*	*	*	*	*	*	*	300	1200	La^{+3}
Sr^{+2}	1·27	10	1000	1000[3]	3000	3000[4]	5000[5]	2000	10	*	20	80	30	100	10	10		100	100	20	2	*	*	30	10	*	*	30	30	1500	1000	Sr^{+2}
Pb^{+2}	1·32	10	*	*	*	*	*	*	*	*	*	*	*	*	*	*		*	*	*	*	*	*	*	*	*	*	*	*	*	10	Pb^{+2}
Ba^{+2}	1·43	5	80	65	200	200	400	600	*	7	10	10	*	60	*	7		10	10	10	*	*	*	20	10	*	*	20	20	*	30	Ba^{+2}
Rb^{+1}	1·49	20	*	*	*	*	*	*	*	*	*	*	*	50	*	*		*	*	*	*	*	*	*	*	*	*	*	*	*	—	Rb^{+1}
Generalized composition from optics or chemical analysis (Wt percentages)			An 65	An 57	An 46	An 41	An 38	An 30	En 42 Fs 17	En 34 Fs 19	En 32 Fs 36	En 23 Fs 50	Fs 61	En 6·5 Fs 68	Fo C.70	Fo 64		Fo 41	Fo 20	Fo 3												

Data for trace elements determined spectrographically except where figure is in brackets when the values given are from chemical analyses.

An asterisk indicates that the element is present in amounts less than the sensitivity.

A dash indicates absence of data.

The following elements were also sought but are not present in amounts above the sensitivity which is indicated in brackets: B(1), Be(10), Ge(10), Sn(5), Zn(300), Sb(300), Ce(500), Cd(300), Ag(1), Tl(50), except for Zn in magnetite, see footnote 2.

r Ionic radius in Ångstrom units at valency shown.

s Sensitivity applying to the plagioclase only. For the other minerals the sensitivity may differ a small amount from that shown.

The results for the various minerals are arranged under columns A to F. All A column minerals are of approximately contemporaneous formation, and so for all B column minerals, etc. The order A, B, C, etc., represents the sequence of differentiation (see page 136).

A Columns—Data for Minerals from Various Early Rocks of the Border Series:—

A I Plagioclase from olivine gabbro 1851 (Perpendicular felspar rock) mean of two determinations (1939, p. 144).

A VII Pyroxene—the major constituents were determined for pyroxene from olivine gabbro 4289 (1939, p. 152) while the trace constituents were determined on essentially similar pyroxene from olivine gabbro 1846 (1939, p. 151).

A XIII Olivine from 1846 (see A VII above).

B Columns—Data for Minerals from Hypersthene Olivine Gabbro and Related Rocks Low in the Accessible Part of the Layered Series:—

B II Plagioclase, B VIII Pyroxene, B XIV Olivine: from hypersthene-olivine gabbro 4077 at 500 m in layered series (1939, p. 89), average of 2 determinations (see Table 1).

B XVIII Ilmenite, B XXIII Magnetite: from melanocratic band 2308 at 850 m in the layered series (1939, p. 92).

C Columns—Data for Minerals from the Middle Gabbros:—

C III Plagioclase, C IX Pyroxene, C XIX Ilmenite, C XXIII Magnetite: from the olivine-free Middle Gabbro 3662 at 1150 m in the layered series (1939, p. 95).

D Columns—Data for Minerals from Hortonolite Ferrogabbro:—

D IV Plagioclase—the major constituents were determined for felspar separated from hortonolite ferrogabbro 2580 at 1700 m in the layered series (300 yds west of Main House, Skaergaard Peninsula) while the trace elements were determined for felspar from 1907 at 1775 m in the layered series (1939, p. 99).

D X Pyroxene, D XV Olivine, D XX Ilmenite, D XXIV Magnetite: from the hortonolite ferrogabbro 1907 at 1775 m in the layered series (1939, p. 99).

D XXXII Apatite—from hortonolite ferrogabbro 3649 at 1875 m in the layered series (1939, p. 104).

E Columns—Data for Minerals from Ferrohortonolite Ferrogabbro:—

E V Plagioclase, E XI Pyroxene, E XVI Olivine, E XXI Ilmenite, E XXV Magnetite: from ferrohortonolite ferrogabbro 4145 at 2175 m in the layered series (1939, p. 104), except that olivine is from 4146 at 2075 m.

F Columns—Data for Minerals from Fayalite Ferrogabbro:—

F VI Plagioclase—from fayalite ferrogabbro 4142 (Purple Band) at 2475 m in the layered series (1939, p. 108).

F XII Pyroxene—from fayalite ferrogabbro 1881 (Purple Band) at 2500 m in layered series (1939, p. 109).

F XVIII Olivine—major constituents determined on olivine from the fayalite ferrogabbro 4139 of the unlaminated layered series at 2600 m (1939, p. 113) while the trace constituents were determined for olivine separated from fayalite ferrogabbro 1881 (see F XII).

F XXVI Magnetite, F XXVIII Apatite: from the fayalite ferrogabbro 4142 (see F VI).

Footnotes.

1. The pyroxenes are dominantly clinopyroxene but some orthopyroxene is present either as the result of exsolution or separation direct from the magma (see p. 150).
2. Magnetites also contain about 1% Zn while Zn in the ilmenites is below the sensitivity which is about 1000 ppm in iron ores. Magnetite from 4142 was not obtained as pure as the others and should only be considered a magnetite concentrate.
3. SrO = 0·24% by chemical methods, *i.e.*, Sr = 2000 ppm.
4. SrO = 0·41% by chemical methods in plagioclase from 2580, *i.e.*, Sr = 3500 ppm.
5. SrO = 0·28% by chemical methods *i.e.* Sr = 2400 ppm.

Table D—Major and trace elements in late granophyre differentiates

	r	s	G	*Hedenbergite Granophyres*				*Transgressive Acid Granophyres*						Z
								Sill			*Veins*			
				I 1225	II 3047	III 3021	IV 1224	V 3058	VI 360	VII 359	VIII 2562	IX 4064	X 3659	
				Major constituents as weight per cent of the oxide										
SiO_2	0·39				58·81			75·03						
Al_2O_3	0·57				12·02			13·17						
TiO_2	0·64				1·26			0·31						
Fe_2O_3	0·67				5·77			1·56						
MgO	0·78				0·72			0·15						
FeO	0·83				9·38			0·58						
Na_2O	0·98				3·91			4·24	4·33		4·58	5·14	4·25	
CaO	1·06				5·03			0·69						
K_2O	1·33				2·39			3·85	3·77		3·44	3·51	4·18	
				Trace constituents as parts per million of the element										
P^{+5}	0·34		(3050)	—	(2800)	—	—	(80)	—	—	—	—	—	(1700)
Ga^{+3}	0·62	1	28	20	50	20	20	20	30	30	25	50	30	25
Cr^{+3}	0·64	1	*	10	*	15	50	5	20	30	25	5	*	33
V^{+3}	0·65	5	*	*	*	*	20	20	25	10	30	20	*	28
Mo^{+4}	0·68	1	*	*	3	*	*	3	1	1	*	*	*	*
Li^{+1}	0·78	1	15	10	30	15	3	5	3	3	30	15	20	16
Ni^{+2}	0·78	2	4	8	8	12	8	5	8	15	8	10	2	6
Co^{+2}	0·82	2	12	15	4	5	*	3	6	2	7	8	*	5
Cu^{+2}	C.0·8	10	500	500	300	400	30	20	*	*	30	20	30	C.12
Sc^{+3}	0·83	10	8	10	*	*	*	*	*	*	*	*	*	*
Zr^{+4}	0·89	10	250	1000	800	2000	1000	1000	500	2000	500	600	500	160
Mn^{+2}	0·91	10	2000	3000	(1500)	3000	800	(80)	80	600	300	—	—	150
Y^{+3}	1·06	30	180	150	200	500	500	200	200	500	200	200	200	C.30
La^{+3}	1·22	30	C.30	30	50	100	300	200	200	500	200	100	200	C.30
Sr^{+2}	1·27	10	400	500	500	600	600	150	200	500	500	400	300	800
Ba^{+2}	1·43	5	400	300	600	1000	2500	1500	800	1000	1500	1500	2000	1800
Rb^{+1}	1·49	20	25	20	40	80	150	300	100	150	150	100	150	160

Data for trace elements determined spectrographically except where figure is in brackets when the values given are from chemical analysis.

An asterisk indicates that the element is present in amounts less than the sensitivity.

A dash indicates absence of data.

The following elements were also sought but are not present in amounts above the sensitivity which is indicated in brackets: B(1), Be(10), Ge(10), Sn(5), Zn(300), Sb(300), Ce(500), Cd(300), Ag(1), Pb(10), Tl(50).

r Ionic radius in Angstrom units at valency shown.

s Sensitivity by the spectrographic method employed.

G Average of two basic hedenbergite granophyres from the top of the layered series (see Table E).

I 1225 Basic augite granophyre transitional to quartz gabbro, Brodretoppen, near 1224 (see below under IV).

II 3047 Hedenbergite granophyre, Brodretoppen (1939, p. 210).

III 3121 Hedenbergite granophyre, Basistoppen (1939, p. 211).

IV 1224 Granophyre, Brodretoppen; related to Hedenbergite granophyres but more acid (1939, pp. 205 and 209).

V 3058 Acid granophyre, sill on Tinden (1939, pp. 204 and 207).

VI 360 Acid granophyre, second specimen from sill on Tinden (1939, p. 204).

VII 359 More melanocratic part of acid granophyre sill 360 (see above under VI).

VIII 2562 " Acid granophyre " vein (1939, p. 206) approaching hornblende porphyrite in composition and texture.

IX 4064 " Acid granophyre " vein (1939, p. 206)—intermediate between VIII and X in mineral composition and texture.

X 3659 " Acid granophyre " vein (1939, p. 206)—a biotite microgranite in mineral composition and texture.

Z Average for the " Grey Gneiss " of the metamorphic complex (see Table E).

Table E—Major and trace elements in certain granophyre inclusions and in the gneisses of the metamorphic complex

	r	s	*Granophyre Inclusions*							*Gneisses of the Metamorphic Complex*											
										"Grey Gneiss"						*"Pink Gneiss"*	*Amphibolite*				
			U	*I* 318	*II* 1836	*III* 1841	*IV* 1840	*V* 1842	*VI* 4074	*VII* 1867	*VIII* 1865	*IX* 4209	*X* 2017	*XI* 341 *A*	*XII* 2802 *A*	*XIII* 2802 *B*	*XIV* 2015	*XV* 341 *B*	*Z*	*V*	
			Major constituents as weight per cent of the oxide																		
SiO_2	0·39									68·17											SiO_2
Al_2O_3	0·57									16·13											Al_2O_3
TiO_2	0·64									0·63											TiO_2
Fe_2O_3	0·67									0·58											Fe_2O_3
MgO	0·78									1·82											MgO
FeO	0·83									2·09											FeO
Na_2O	0·98			5·35	6·14	5·45	5·08	—	—	4·40	3·98										Na_2O_3
CaO	1·06									2·07											CaO
K_2O	1·33			2·62	1·50	2·40	1·02	—	—	3·22	2·93										K_2O
			Trace constituents as parts per million of the element																		
P			—	—	—	—	—	—	—	(1700)	—	—	—	—	—	—	—	—	(1700)	—	P
Ga^{+3}	0·62	1	**40**	20	50	50	30	40	40	20	20	25	20	35	30	20	25	15	**25**	**24**	Ga^{+3}
Cr^{+3}	0·64	1	**18**	*	*	*	60	20	30	10	15	3	20	100	50	20	50	600	**33**	**85**	Cr^{+3}
V^{+3}	0·65	5	**13**	20	*	*	20	20	15	20	50	20	20	20	40	10	150	150	**28**	**50**	V^{+3}
Mo^{+4}	0·68	1	*	*	*	*	*	*	*	*	*	*	*	*	*	*	*	*	*	*	Mo^{+4}
Li^{+1}	0·78	1	**2**	2	3	3	2	1	1	20	25	15	10	10	20	1	25	15	**17**	**16**	Li^{+1}
Ni^{+2}	0·78	2	**12**	7	20	20	2	5	15	4	1	3	2	20	3	*	60	150	**6**	**24**	Ni^{+2}
Co^{+2}	0·82	2	**3**	5	**3**	5	*	*	3	6	5	5	*	8	7	*	30	35	**5**	**11**	Co^{+2}
Cu^{+2}		10	**20**	20	20	10	*	*	60	*	*	*	30	30	*	*	20	20	**C.15**	**C.15**	Cu^{+2}
Sc^{+3}	0·83	10	*	*	*	10	*	*	*	*	*	*	*	*	10	*	10	30	*	*	Sc^{+3}
Zr^{+4}	0·87	10	**2500**	500	+1%	3000	200	500	2000	250	150	200	100	200	70	20	150	30	**160**	**140**	Zr^{+4}
Mn^{+2}	0·91	10	**300**	—	—	—	300	300	300	(350)	—	—	100	80	400	20	800	2000	**150**	**400**	Mn^{+2}
Y^{+3}	1·06	30	**350**	200	200	300	300	300	800	*	*	*	30	*	50	*	60	50	**C.30**	**C.30**	Y^{+3}
La^{+3}	1·22	30	**280**	300	200	300	100	300	500	30	30	30	*	30	50	*	30	*	**C.30**	**C.30**	La^{+3}
Sr^{+2}	1·27	10	**130**	100	70	200	20	100	300	2000	1000	500	800	600	600	800	400	200	**800**	**700**	Sr^{+2}
Pb^{+2}	1·32	10	*	*	*	*	*	*	*	*	*	*	*	*	15	10	*	*	*	*	Pb^{+2}
Ba^{+2}	1·43	5	**350**	500	10	400	15	20	1200	3000	2000	2000	3000	300	600	1500	400	30	**1800**	**1500**	Ba^{+2}
Rb^{+1}	1·49	20	**30**	30	*	30	*	*	80	250	150	100	150	30	250	300	80	*	**160**	**150**	Rb^{+1}

Data for trace elements determined spectrographically except where figure is in brackets when the values given are from chemical analysis.
An asterisk indicates that the element is present in amounts less than the sensitivity.
A dash indicates absence of data.
The following elements were also sought but are not present in amounts above the sensitivity which is indicated in brackets: B(1), Be(10), Ge(10), Sn(5), Zn(300), Sb(300), Ce(500), Cd(300), Ag(1), Tl(50).
r Ionic radius in Ångstrom units at valency shown.
s Sensitivity by the spectrographic method employed.

U Average composition of granophyre inclusions I–VI.
I 318, II 1836, III 1841, IV 1840 — Granophyre inclusions in the western border group. For localities see 1939, p. 187.
V 1842 Granophyre inclusion similar to the above from western border group collected near 1840.
VI 4074 Granophyre inclusion from low in the layered series (1939, p. 185).
VII 1867, VIII 1865 — "Grey gneiss", Mellemo, Skaergaard area (1939, p. 11).
IX 4209 "Grey gneiss", 2 km NNW of Bagnaesset, Kangerdlugssuaq, East Greenland.
X 2017 "Grey gneiss" north side of Hutchinson glacier, 28 km from snout.
XI 341A At 1930 depot on coast ½ km south of end of Kangerdlugssuaq Glacier, Kangerdlugssuaq, East Greenland.
XII 2802 A "Grey gneiss" west point of Hovedvejs Nunatak, 20 km west of inner Kangerdlugssuaq, East Greenland.
XIII 2802 B "Pink gneiss", the pegmatitic part of 2802 A (see XII above).
XIV 2015 Acid amphibolite of quartz-diorite composition, ¾ mile east of 2017 (see X above).
XV 341 B Basic augite bearing amphibolite adjacent to 341A (see XI above).
Z Average content of trace constituents in "Grey gneiss"—mean of VII, VIII, etc., to XII.
V Rough estimate of content of trace constituents in the metamorphic complex (11 parts of "Grey gneiss" column V; 3 parts of average amphibolite; and 1 part of "Pink gneiss", column XIII).

Table F—Average values to illustrate the general trend in amounts of major and trace elements in the differentiates of the Skaergaard intrusion. (Sequence of differentiates A–I)

	r	s	Y	*Early Border Group* A	A¹	*Layered Series* B	C	D	E	F	G	*Transgressive Granophyres* H	I	Z	
Major constituents as weight per cent of the oxide															
SiO_2	0·39		47·92	41·27	46·15	46·37	46·90	44·81	44·61	41·13	52·13	58·81	75·03	68·17	SiO_2
Al_2O_3	0·57		18·87	8·71	21·34	16·82	16·55	13·96	11·70	17·88	15·87	12·02	13·17	16·13	Al_2O_3
TiO_2	0·64		1·40	1·54	0·55	0·79	2·61	2·55	2·43	2·48	1·14	1·26	0·31	0·63	TiO_2
Fe_2O_3	0·67		1·18	2·69	0·79	1·52	2·96	3·75	2·05	4·05	5·61	5·77	1·56	0·58	Fe_2O_3
MgO	0·78		7·82	27·09	9·97	9·61	5·80	5·54	1·71	·25	1·11	0·72	0·15	1·82	MgO
FeO	0·83		8·65	10·52	8·21	10·44	12·18	16·66	22·68	26·63	11·17	9·38	·58	2·09	FeO
Na_2O	0·98		2·44	0·69	1·72	2·45	2·97	3·35	2·95	2·15	3·63	3·91	4·24	4·40	Na_2O
CaO	1·06		10·46	6·59	10·24	11·29	9·67	8·53	8·71	10·03	5·80	5·03	0·69	2·07	CaO
K_2O	1·33		0·19	0·13	0·12	0·20	0·21	0·33	0·35	0·47	1·38	2·39	3·85	3·22	K_2O
Trace constituents as parts per million of the element															
P^{+5}	0·34		(280)	(90)	(220)	(260)	(290)	(350)	(8050)	(7000)	(3050)	(2800)	(80)	(1700)	P^{+5}
Ga^{+3}	0·62	1	17	8	15	19	15	20	20	17	28	35	30	25	Ga^{+3}
Cr^{+3}	0·64	1	170	1500	300	230	*	*	*	1	*	8	14	33	Cr^{+3}
V^{+3}	0·65	5	140	120	180	220	400	13	*	*	*	*	18	28	V^{+3}
Mo^{+4}	0·68	1	*	3	*	*	*	*	*	*	*	2	1	*	Mo^{+4}
Li^{+1}	0·78	1	2	2	2	2	3	2	2	6	15	25	13	17	Li^{+1}
Ni^{+2}	0·78	2	170	1000	350	120	40	*	*	*	4	10	8	6	Ni^{+2}
Co^{+2}	0·82	2	53	90	80	48	40	30	20	7	12	5	5	5	Co^{+2}
Cu^{+2}	C.0·8	10	130	100	45	67	180	350	600	300	500	350	20	C.15	Cu^{+2}
Sc^{+3}	0·83	10	12	*	10	20	20	10	15	*	8	*	*	*	Sc^{+3}
Zr^{+4}	0·89	10	50	30	30	33	30	15	20	25	250	1400	850	160	Zr^{+4}
Mn^{+2}	0·91	10	(800)	(1200)	(1100)	(700)	(1050)	(1300)	(1620)	(3700)	2000	(1500)	C.150	150	Mn^{+2}
Y^{+3}	1·06	30	*	*	*	*	*	*	*	70	180	350	250	C.30	Y^{+3}
La^{+3}	1·22	30	*	*	*	*	*	*	*	*	C.30	80	240	C.30	La^{+3}
Sr^{+2}	1·27	10	350	100	300	600	500	500	500	300	400	550	350	800	Sr^{+2}
Ba^{+2}	1·43	5	43	10	15	18	50	45	50	75	400	800	1400	1800	Ba^{+2}
Rb^{+1}	1·49	20	*	*	*	*	*	*	*	15	25	60	160	160	Rb^{+1}
S	1·74		(500)	—	—	(450)	(400)	—	(600)	(4100)	(450)	—	—	—	S

The data for spectographically determined elements are averages taken from Tables A, B, D and E.

The data for chemically determined constituents in some cases are average values, but in others only one analysis has been made (see also Tables A, B, D and E).

Data for trace elements were determined spectrographically except where figure is in brackets when the values given are from chemical analysis.

An asterisk indicates that the element was present on the average in amounts less than the sensitivity.

A dash indicates absence of data.

The following elements were also sought but are not present, on the average, in amounts above the sensitivity which is indicated in brackets: B(1), Be(10), Ge(10), Sn(5), Zn(300), Sb(300), Ce(500), Cd(300), Ag(1), Pb(10), Tl(50).

r Ionic radius in Ångstrom units at valency shown.

s Sensitivity by the spectrographic method employed.

Y Average composition of fine grained marginal gabbro taken to represent the composition of the original Skaergaard magma (Table A, column Y).

A Gabbro picrite, Northern border group (Average of columns I and II, Table C).

A¹ Earliest olivine gabbro (perpendicular felspar rock, etc.), outer part of the Western border group (Average of columns III and IV, Table C).

B Hypersthene-olivine gabbro (Average of one part of column VII A and two parts of column VII B, Table C).

C " Middle gabbro "—olivine free (Average of columns VIII and IX, Table C).

D Hortonolite ferrogabbro (Average of columns X and XI, Table C).

E Ferro-hortonolite ferrogabbro column XII, Table C.

F Fayalite ferrogabbro (Average of columns XIII–XV, Table C).

G Basic hedenbergite granophyre of layered series (Average of columns XVIII and XIX, Table C).

H Transgressive hedenbergite granophyre (Average of columns II and III, Table D).

I Average acid granophyre (Average of columns V to X, Table D).

Z Average " Grey gneiss " of the metamorphic complex (Table E, column Z).

BOOK REVIEW

The Atmospheres of the Earth and Planets (Edited by GERARD P. KUIPER). University of Chicago Press, Chicago. 1949. $7.50.

The book comprises a collection of papers presented in September, 1947 by astronomers, geologists, geochemists, meteorologists and physicists at a Symposium arranged by the University of Chicago. Before going to press in 1948 the manuscripts were revised and more recent work incorporated. In a subject whose development has become so rapid no book can be quite up-to-date but the present volume contains so much fundamental material that it is certain to retain its value as a source of reference for many years to come. To several of the topics treated here, useful supplements may be found in reports on the more recent discussion on planetary atmospheres held at Dublin under the auspices of the *Royal Astronomical Society* a few months ago.

The great wealth of the material contributed by eighteen experts is such that in the space available no detailed review is possible. The best way of indicating the range covered by this survey seems to be to give here a complete list of the papers it contains:

GERARD P. KUIPER, Introduction; CARL-GUSTAF ROSSBY On the Nature of the General Circulation of the Lower Atmosphere; H. C. VAN DE HULST Scattering in the Atmospheres of the Earth and the Planets. Three papers on the Upper Atmosphere Studied from Rockets: JESSE L. GREENSTEIN Research Programs; H. E. CLEARMAN The Ultraviolet Solar Spectrum; E. DURAND Rocket Sonde Research at the Naval Research Laboratory. FRED L. WHIPPLE, LUIGI JACCHIA and ZDENEK KOPAL Seasonal Variations in the Density of the Upper Atmosphere; P. SWINGS The Spectra of the Night Sky and the Aurora; LYMAN SPITZER, Jr. The Terrestrial Atmosphere above 300 km; ROLLIN T. CHAMBERLIN Geological Evidence on the Evolution of the Earth's Atmosphere; HARRISON BROWN Rare Gases and the Formation of the Earth's Atmosphere; ARTHUR ADEL Selected Topics in the Infra-red Spectroscopy of the Solar System; M. V. MIGEOTTE On the Presence of CH_4, N_2O, and NH_3 in the Earth's Atmosphere; THEODORE DUNHAM, Jr. Spectroscopic Observations of the Planets at Mount Wilson; GERARD P. KUIPER Survey of Planetary Atmospheres; G. HERZBERG Laboratory Absorption Spectra Obtained with Long Paths; JAMES FRANCK Possibility of Photosynthesis on Mars.

The book is excellently produced. The great number of diagrams and of plates (representing mostly solar, lunar and planetary spectra) accounts for its fairly high price. F. A. PANETH (Durham)

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